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Plio-Pleistocene sedimentation in West Turkana (Turkana Depression, Kenya, East African Rift System): paleolake fluctuations, paleolandscapes and controlling factors

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ABSTRACT

Pliocene and Pleistocene sediments from West Turkana (Kenya, East African Rift System) form emblematic syn-rift successions for understanding the evolution of extensional basin and continental rifting. They also constitute world-renowned fossil-bearing strata from which >500 hominin fossils were discovered over the past decades, with >100 inventoried archaeological sites. However, associated sedimentary dynamics and architectures as well as paleoenvironments are only partially reconstructed and the relative contribution of climate

and tectonism to paleoenvironmental change over time remains unclear. Here, through the interpretation of sedimentary facies, the delineation of sequences and the analysis of $\delta^{13}C$ in soil carbonates, we provide the first exhaustive reconstruction from ~4 to ~0.75 Ma of (i) fluctuations of the paleolake that occupied the Turkana Depression, (ii) the successive sedimentary dynamics and related paleolandscapes that characterized the West Turkana area and (iii) respective roles of climate and tectonism in the sedimentation. We show evidence for seven major transgression-regression (T-R) cycles between ~4 and ~0.75 Ma superimposed locally by lower amplitude T-R cycles. , Comparing the sedimentological interpretations and the $\delta^{13}C$ values in soil carbonates (literature data), we reveal that fluctuations of rainfall over the Ethiopian Dome, which hosts the drainage basin of the Omo River — the main tributary of Lake Turkana — controlled high-amplitude lake level fluctuations during the Plio-Pleistocene period. We also demonstrate that vegetation and tree cover evolved differently in the Omo Valley and West Turkana. Furthermore we outline that two different sedimentary systems reflecting two distinct modes of sedimentation alternated through times in the West Turkana area as a response to the variations in sediment supply coming from the western rift shoulder (i.e. Lapurr Range) that alternatively generated wave- or river-dominated sedimentary systems. In conclusion, we reveal that climate regulated water input, paleolake water-level fluctuations and vegetation. Tectonism determined sediment supply to the basin controlled in West Turkana by pulses of increased activity of the main border fault (i.e., the Murua Rith Lapurr Fault).

KEYWORDS: Nachukui Fm, Omo Group, rift basin, Pliocene

1. INTRODUCTION

The Turkana Depression has been investigated for several decades, the deposition and preservation of thick syn-rift strata in this area providing a thorough record of the evolution of

the East African Cenozoic rifting from ~35 million years ago to the present. In the early1980's, ambitious seismic imagery campaigns (Amoco Kenya Petroleum company and project PROBE) boosted geological investigations, while the first detailed stratigraphical and sedimentological analyses were performed (de Heinzelin, 1983; Brown and Feibel, 1986; Harris *et al.*, 1988). Seismic imagery revealed that the Turkana Depression is composed of several individual grabens and half-grabens filled by >3 km of Neogene sediments (Dunkelman *et al.*, 1988; 1989; Morley *et al.*, 1992). After the 1990's, additional seismic reflection, gravity and magnetic data, deep and shallow wells and geological analysis of outcrops led to the elaboration of a complex structural framework that shed light on fundamental geotectonic processes associated with continental rifting (Morley *et al.*, 1999; Vétel and Le Gall, 2006; Brune *et al.*, 2017; Ragon *et al.*, 2019; Morley, 2020). In parallel, sedimentary processes in the Turkana Depression (Cohen 1981; Frostick *et al.*, 1986; Williamson and Savage, 1986; Frostick and Reid, 1987; Tiercelin *et al.*, 2010; Hargrave *et al.*, 2014; Nutz *et al.*, 2017; Schuster and Nutz, 2018) stand as text-book examples of sedimentation in extensional basins, with thick exposures of several hundred meters that provide a quasi-complete record of extensional basin evolution, especially during the Plio-Pleistocene, between ~4.00 and ~0.75 million years (Ma). Well-exposed in distinct parts of the northern region of the Turkana Depression, the three formations of the Omo Group (de Heinzelin, 1983), i.e. the Shungura Fm, the Koobi Fora Fm and the Nachukui Fm, constitute exceptionally continuous archives from the three key domains of a half-graben system, i.e. the axial system, the ramp transverse system, and the border-faulted transverse system, respectively.

The three formations are also world-renowned fossil-bearing successions for paleoanthropology, paleontology and archeology. Dedicated investigations are ongoing since Arambourg's expeditions in the region (Arambourg 1935 and 1943) some 30 years after the

trans-African expedition of du Bourg de Bozas that reported fossil vertebrates from the Lower Omo Valley (Boisserie *et al*., 2008).The Nachukui Fm happened to be particularly rich with, to date, >100 archaeological sites and >500 specimens of early hominins, some of which have been instrumental for the understanding of human evolution. These discoveries include specimens of *Australopithecus anamensis, Kenyanthropus platyops, Paranthropus aethiopicus, Paranthropus boisei, Homo rudolfensis, Homo erectus* and *Homo sapiens* (Brown *et al.*, 1985; Walker *et al.*, 1986; Leakey and Walker, 1988; Leakey *et al.*, 1995, 2001; Prat *et al.*, 2003, 2005). Numerous stone tools were also found in strata of this Formation (Roche *et al*., 1999; Delagnes and Roche, 2005), among which the earliest evidence of Acheulean stone tool technology (Lepre *et al.*, 2011). More recently, the Lomekwian stone tools found in strata of the Nachukui Fm pushed the first appearance of homonin technological behaviour back to 3.3 Ma (Harmand *et al.,* 2015). In summary, the northern Turkana Depression is generally considered a key area to understand how paleoenvironmental change and human evolution are linked (Maslin and Trauth, 2009; Maslin *et al.*, 2014; Potts and Faith, 2015; Cohen *et al.*, 2016). For all these reasons, the Plio-Pleistocene sediments of the Turkana Depression, and particularly the Nachukui Fm, are of uttermost importance for many communities within the geosciences.

However, since the pioneer work by Harris *et al*. (1988) on the Nachukui Fm, by de Heinzelin (1983) on the Shungura Fm, and by Brown and Feibel (1991) on the Koobi Fora Fm, none of the three formations of the Omo Group has been comprehensively investigated. Instead, a variety of local studies have been performed on each Formation, often in direct association with paleontological and archeological discoveries. In this review, we present a synthesis of recent investigations on the Nachukui Fm, which is the longest-lasting Formation of the Omo Group, and we refine conditions of sedimentation between ~4.00 and ~0.75 Ma, in West Turkana and subsequently paleoenvironmental changes. For the first time, continuous

and comprehensive reconstructions of i) paleolake fluctuation, ii) the evolution of sedimentary dynamics and related paleolandscapes and iii) the respective influence of various controlling factors (i.e., climate, tectonism) are proposed based on depositional environments analysis, sequence stratigraphy and paleovegetation in both the Omo Valley and West Turkana areas. Successive sedimentary systems are characterized revealing that variable sedimentary paleolandscapes alternated along the border fault (i.e., Murua Rith Lapurr Fault or MRLF) of the northern Turkana Depression between ~4.00 and ~0.75 Ma. Confronting $\delta^{13}C$ isotopic ratios in soil carbonates (Levin, 2015) and inferred tree cover (Cerling *et al.*, 2011) with our sedimentological interpretations; we unravel the relative influence of climate and tectonism on paleolake fluctuations and paleolandscape evolution. In providing this synthesis we aim at documenting depositional processes in continental rift basin, especially along faulted borders. We also aim at proposing a general framework of paleoenvironmental reconstruction for further investigations in this area.

2. GEOLOGICAL CONTEXT

2.1. Regional structural setting

The Turkana Depression is part of the East African Rift System (EARS), located between the Ethiopian and the East-African Domes (Fig. 1A). It corresponds to a 300 km long and 200 km wide lowland that separates the Main Ethiopian Rift to the north from the northern Kenyan rift to the south (Ebinger *et al.*, 2000; Brune *et al.,* 2017). The Turkana Depression consists of several N-S oriented half-grabens and grabens which developed from the middle to upper Eocene up to the present-day (Rosendahl, 1987; Dunkelman et al., 1988, 1989; Morley *et al.*, 1999; Tiercelin and Lezzar, 2002; Vétel and Le Gall, 2006; Tiercelin *et al.*, 2012a and 2012b; Macgregor, 2015; Boone *et al.*, 2017, 2019; Ragon *et al.*, 2019).

The Turkana Depression is subdivided into a southern and a northern depression based on their different structural characteristics (Ragon *et al*., 2019). The southern Turkana Depression consists of five juxtaposed N-S oriented half-grabens known as the Naipa, Lokichar, Turkwel, Kerio and South Lake Basins (Fig.1B, C). Together they form a > 200 km wide rift system (*sensu* Buck, 1991), characterized by relatively low-lying reliefs due to limited shoulder uplifts in between the different basins. The onset of the extension related to the Cenozoic EARS started at around 45-40 Ma in the Naipa and Lokichar Basins (Boone *et al*., 2019). During the early Miocene (23-15 Ma), surface rupture migrated farther east with the opening of the Turkwel and Kerio Basins (Vétel and Le Gall, 2006). Later, the initiation of the South Lake Basin may have started slightly before 10 Ma (Morley *et al.*, 1999; Vétel and Le Gall, 2006). In contrasts, the northern Turkana Depression is strikingly different. It consists of a single 80 km wide and >4 km deep (Abdelfettah *et al.*, 2016) N-S oriented half-graben (Hendrie *et al.*, 1994; Morley *et al.*, 1999; Tiercelin *et al.*, 2004; Vétel and Le Gall, 2006; Ragon *et al*., 2019) referred to as the North Lake Basin (Fig. 1B, C). The North Lake Basin forms a narrow rift (*sensu* Buck, 1991) bounded to the west by the N-S oriented Murua Rith-Lapurr Fault (MRLF) (Fig. 1C, 2), and to the east by a west-dipping ramp system. West of the MRLF, a relative high-relief characterizes the rift shoulder referred to as the Murua Rith Hills and the Lapurr Range (Fig. 2). East of the MRLF, the basin is affected by several smaller normal faults up to a second major one, i.e. the Shore Fault (Nutz *et al.*, 2017; Ragon *et al*., 2019), as it is currently located at Lake Turkana's shore (Fig.2). In the northern Turkana Depression, the onset of the extension is delayed. It is associated with the development of micro-basins dated between ~28 and ~25.5 Ma (Ragon *et al.*, 2019), during the first phase of extension attributed to EARS1 (*sensu* Macgregor, 2015). After a period of tectonic quiescence, a second pulse of extension referred to as EARS2 (*sensu* Macgregor, 2015), led to the opening of the North Lake Basin around 14 Ma (Boone *et al.*, 2017; Ragon *et al.*, 2019)

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and the onset of sedimentation in the basin. At ~0.7 Ma, the shore fault (Fig. 2) became much more active than the MRLF, the depocentre shifted markedly eastward (Nutz *et al.*, 2017) and this event marked the end, except locally, of the Nachukui Fm sedimentary system. The deposits of the Nachukui Fm subsequently constituted the footwall of the newly-formed and still active depocentre (Nutz *et al.*, 2017), providing geologists with some of the remarkable outcrops presented hereafter.

2.2. Regional lithological framework

The basement of the northern Turkana Depression is made of metamorphic Proterozoic rocks (Williamson and Savage, 1986; Morley *et al.*, 1992; Wescott *et al.*, 1993; Morley *et al.*, 1999; Tiercelin *et al.*, 2012b). The oldest sedimentary rocks deposited on the basement are the Turkana Grits, locally known as the Lapur Sandstone Formation, which correspond to fluvial sandstones from Late Cretaceous to Oligocene age (Murray-Hughes, 1933; Arambourg and Wolff, 1969; Morley *et al.*, 1999; Tiercelin *et al.*, 2004, 2012a and 2012b). Overlying the Turkana Grits, thick lava flows commonly associated with intercalated volcano-sedimentary rocks were deposited as a response to intense fissural volcanic activity during Late Eocene to mid Miocene times, depending of the location. They are referred to as the Turkana Volcanics Fm (Fig. 3; Bellieni *et al.*, 1981, 1987; Zanettin *et al.*, 1983; McDougall and Brown, 2009; Rooney, 2017). The 3 km thick Turkana Volcanics Fm is spread over an extensive region, including the West Turkana area (Fig. 3). In West Turkana, the Turkana Volcanics Fm is Upper Eocene to Lower Oligocene in age and predates rifting (Ragon *et al.*, 2019). Overlying the Turkana Volcanics Fm, Oligocene sedimentary rocks were deposited in extensional micro-basins (Ragon *et al.*, 2019); they fill the oldest syn-rift structures in this area. Upwards in the succession, Miocene rocks are exposed sparsely, but they have been identified in some wells (Schofield *et al.,* 2020) suggesting deposition at least in the North Lake Basin after its

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opening 14 Ma ago. Overlying, Plio-Pleistocene sediments of the Omo Group are exposed all around modern Lake Turkana. The Omo Group, first defined by de Heinzelin (1983), is subdivided into three main formations referred to as the Nachukui, Koobi Fora and Shungura fms cropping out in the western Turkana, eastern Turkana and Omo Valley, respectively (Fig. 1B), and two additional minor ones referred to as Usno and Mursi Fms both exposed in the Omo Valley. Younger sediments are included in the Turkana Group (Feibel, 2011), such as the Kibish Fm (Butzer, 1971) and the Holocene Galana Boi Fm (Owen and Renaut, 1986), which unconformably covers sediments of the Omo Group (Fig. 3).

The Nachukui Formation is located in the West Turkana area and represents a >700 m thick succession of fossil-bearing fluvial-deltaic-lacustrine sediments (Harris *et al.*, 1988; Feibel, 2011). This formation is divided into 8 members separated by tuff horizons. It spans a time interval ranging from ~4.00 to ~0.75 Ma based on tephrostratigraphy (Brown *et al.*, 2006; McDougall *et al.*, 2012). Exposures are located eastward of the MRLF and westward of the Shore Fault (Fig. 2).

During the Plio-Pleistocene, several volcanic events occurred in the northern Turkana Depression and episodic volcanic material intercalated in sediments of the Nachukui Fm. The Gombe Basalt, which is made of extensive and thick basaltic lava flows corresponds to an important volcanic event that occurred between 4.29 and 4.18 Ma in the north (Erbello and Kidane, 2018) and that persisted southward at least until 3.95 Ma (Haileab *et al.*, 2004; Bruhn *et al.*, 2011). Less widespread, the Kulal Volcanics (dated from ~3.00 to ~2.20 Ma) and the Lenderit Basalt (dated from ~2.20 to ~2.00 Ma) were emplaced mostly in the southeastern part of the northern Turkana depression (Bruhn *et al.*, 2011). Postdating the Nachukui Fm, other volcanic events are at the origin of the North, Central and South islands of Lake Turkana (Brown and Carmichael, 1971; Karson and Curtis, 1994).

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3. METHODS

3.1. Field analysis

Four field surveys were conducted in July 2014, November 2014, July 2015 and July 2016. Eighteen composite sections were measured along river incisions (Fig. 4) and multiple panoramas were analyzed (see Table S1 for detailed locations). We logged sections spanning the time interval between ~4.00 and ~1.25 Ma. The interval between ~1.25 and ~0.75 Ma was not logged by authors but we used information provided by Harris *et al.*, (1988). Sections were measured using a Jacob's staff and we consider a +/- 10% potential error about strata thicknesses. Sedimentary facies were interpreted based on macroscopic examinations during field surveys. Lithology, grain size, sorting, bed thickness, sedimentary structures, and paleocurrents were assessed based on conventional facies analysis (Table 1). Measured sections were correlated based on tephrochronology (Table 2) or through the physical tracing of prominent stratigraphical surfaces within the basin. Subsequently, sequence analysis was carried out. Sequence stratigraphy aims at deciphering cyclicity in sedimentation due to variations in base level (i.e., lake level in this case), flux of sediment supply and/or subsidence rate (Catuneanu, 2018). In this study, we use the model of Trangressive-Regressive sequence (Johnson and Murphy, 1984; Embry and Johannessen, 1992) in order to reconstruct trajectory of the shoreline through time and thus paleolake extensions and contractions. Transgressive-Regressive sequences (T-R sequence) were delineated by considering each sequence as a full cycle of change in the accommodation space, involving an increase due to the landward migrating shoreline during lake level raise followed by a decrease in the space available for sediments to fill associated with the basinward migration of the shoreline due either to lake level fall or sediment wedges progradations. In this model, a sequence is bounded by maximum regressive surfaces that form the Sequence Boundary (SB). The transgressive part includes sediments deposited during the transgression while the regressive part comprises

sediments deposited during the highstand normal regression, the forced regression and the lowstand normal regression (Catuneanu, 2018). The T-R sequence model reflects successive lateral shoreline migrations, landward during the transgressive part and basinward during the regressive part of the T-R sequences. As a consequence, this model highlights paleolake extensions/contractions rather than changes in water depth. In places, several remarkable stratigraphic intervals or surfaces are interpreted such as Maximum Flooding Intervals (MFI) and Wave Ravinement Surfaces (WRS) according to Catuneanu (2018)..

3.2. Chronological constraints

Our chronological framework is established based on tephrochronology using a dataset of 21 dated volcanic tuffs, many of which have been identified throughout the area from their geochemical signatures. Tuffs were collected during logging (Table 2). The chemical composition of 18 tephra layers was analyzed using a microprobe (CAMECA SX-FIVE, CAMPARIS service, Paris 6 University) to measure abundance of Fe_2O_3, CaO and K_2O oxides and Mn and Ti elements. Three tuffs were not analyzed as they had been identified previously by Lepre *et al.*, (2011) and Harmand *et al.*, (2015). Geochemical signatures of these tuffs were compared to published data for identifications (Harris *et al.*, 1988; Brown *et al.*, 2006; McDougall and Brown, 2008). The 18 tephra layers recognized in our sections were identified as 16 distinct tuffs, for which absolute ages estimated by K-Ar and Ar-Ar methods were then obtained from the literature (Pickford *et al.*, 1991; Lepre *et al.*, 2011; McDougall *et al.*, 2012; Harmand *et al.*, 2015; Boës *et al.*, 2018).

3.3. Pedogenic $\delta^{13}C$ record and inferred paleovegetation changes

Plant fossils are rare in Formations of the Turkana Depression, but an important dataset of carbon isotopic measurements was acquired over the years (Cerling, 1999; Harris and Leakey,

2003; Wynn, 2004; Bobe *et al*., 2007; Levin *et al.*, 2011) and recently compiled by Levin (2015). The $^{12}C/^{13}C$ ratio measured in soil carbonates (or $\delta^{13}C_{soil\ carbonate}$) reflects the proportion of plant biomass using C_3 or C_4 photosynthesis pathways (Cerling and Quade, 1993). In the tropics, C_3 plants are predominantly trees and shrubs (and aquatic and montane grasses in a lower proportion), while C_4 plants are predominantly lowland grasses, sedges, and the often overlooked xerophytic and salt-loving shrubs and forbs of the Amaranthaceae family (Tieszen *et al.*, 1979; Sage, 2004). $\delta^{13}C$ in soil carbonates, therefore, has been used to approximate the prevalence of woodlands and grasslands, and to quantify changes in the proportion of woody cover through time. $\delta^{13}C_{soil\ carbonate}$ values in the range -15‰ to -7‰ indicate vegetation with >40% woody cover (such as forest and undifferentiated woodland, shrubland or bushland) and values >-7‰ indicate more open vegetation (such as wooded grasslands and grasslands) (Cerling *et al.*, 2011).

Woody cover can in turn be used as a proxy of climate as more rainfall and/or shorter dry seasons favor denser woody cover (Staver *et al.*, 2011). In the arid Turkana Depression, dense tree cover predominantly occurs in the alluvial plain along the Omo River, where a shallow water table sustains perennial riparian woodlands and forests (Carr, 1998). Considering that the amount of groundwater available in alluvial plains is controlled by the activity of rivers, themselves dependent of rainfall over river catchments, wetter periods should be associated with more important woody cover (forest, woodlands) in opposition to drier periods which should be characterized by less abundant woody cover (grasslands or steppes). To approximate climate change, namely change in rainfall over river catchments, we therefore used $\delta^{13}C$ in soil carbonates. We considered altogether the paleosoil data from Nachukui Fm (n=130) to illustrate paleovegetation changes in the West Turkana area and paleohydrological fluctuations in river catchments of the western rift shoulder (between 3.9 and 0.9 Ma). Data from the Shungura Fm (n=49) were considered independently to illustrate

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paleovegetation and paleohydrological changes in the Ethiopian Highlands and the Omo River catchment (between 3.20 and 1.25 Ma). Subsequently, we compared the reconstructed paleolake extension with woody cover changes in the both West Turkana and Omo Valley areas. In case of matching between periods of lake highstand and important woody cover or conversely between periods of lake lowstand and low woody cover, lake highstands and lowstands are coeval with wet and dry periods in river catchments, respectively, revealing that climate is the dominant parameter controlling water-level fluctuations. In contrast, in case of a lake highstand occurring contemporaneously with a reduced woody cover in both Omo Valley and West Turkana, climate in the considered region can be discarded to explain paleolake fluctuations.

To objectively identify whether significant changes in $\delta^{13}C$ occurred and if so, when, we have developed a new automatic classification method based on *k*-means clustering (Bell, 1978; Grim, 1987; Wagstaff *et al.*, 2001). *K*-means clustering aims at partitioning *n* observations into *k* clusters in which each observation belongs to the nearest cluster. Clusters are selected as such to minimize the variance within clusters (group highly similar observations) and to maximize variation among clusters (differentiate distinct groups). Applying this method to time series data, temporal overlap was not permitted among clusters and each cluster represents a particular period bounded by the oldest and the youngest observations within the cluster. A centroid is estimated for each cluster, defined by the mean age and the mean value of $\delta^{13}C$ of all observations included in a given cluster; centroids are linked forming curves that reflect changes of $\delta^{13}C$ isotopic ratios in soil carbonates through time. The *k*-means method usually requires the number of clusters to be arbitrarily selected prior to analysis, for example using Elbow methods (Bholowalia *et al.*, 2014) or principal component analysis (PCA). For this study, we did not arbitrarily set the number of groups *a priori* to avoid subjective association of observations into artificial time slices. Instead, we ran

12 classifications with a number of clusters ranging from 4 to 15. Considering the 12 classifications, we calculated a total of 114 centroids. In each classification, centroids of groups including at least 3% of the observations were considered representative. Representative centroids were then stacked and linked to reconstruct general changes in the $\delta^{13}C$ isotopic ratios in soil carbonates through time. The confidence interval for each centroid is represented by an envelope that depends on the total variance within the cluster divided by the number of observations included in the cluster.

3.4. Accommodation and shoreline trajectory

As defined by Catuneanu (2018), accommodation (A) is the space made available for potential sediment accumulation. It derives from sea/lake-level changes and rates ofsubsidence/uplift or a combination of these processes. Variations in accommodation during a given time interval (δA) result from changes in base-level combined with vertical movements of the basin substratum (i.e., subsidence) during this time interval. The ratio of accommodation rate/rate of sediment supply or δA/δS ratio is considered to directly drive the trajectory of the shoreline. When $\delta A/\delta S \leq 0$, erosion occurs, progradation (i.e., basinward migration of the shoreline) occurs when $0<\delta A/\delta S<1$ (i.e., δS exceeds δA), aggradation (i.e., stabilization of the shoreline) when $\delta A/\delta S=1$ and retrogradation (i.e., landward migration of the the shoreline) when $\delta A/\delta S>1$ (i.e., δA exceeds δS). Hence, with a constant rate of sediment supply (δS), variation in accommodation space (δA) controls the establishment and rates of progradation and retrogradation dynamics and the corresponding regressive and transgressive trends, respectively. The higher δS compared to δA, the faster is the progradation, the higher δA compared to δS, the faster is the retrogradation. Conversely, with constant rate of variation in accommodation (δA), variation in sediment supply (δS) drives

the establishment and rates of progradation and retrogradation dynamics and the corresponding regressive and transgressive trends.

On a vertical section, variations in accommodation during a given time interval is quantified by summing up the decompacted thickness of sediments (according to Sclater and Christie, 1980) deposited during this considered time interval and the variations of paleobathymetry/paleoaltitude from T_0 to T_1 (Jervey, 1988). Hence, evaluating variations in accommodation requires (1) absolute datums across the section, (2) lithological data and (3) estimations of paleobathymetry/paleoaltitude along each datums. The identification of tuffs in sections of the Nachukui Fm provides chronologically-constrained sediment intervals. The averaged lithology of each interval (i.e. percentages of sand and mud between two datums) is obtained from sedimentological analysis of the sections (Fig. 4) and thicknesses are then corrected for compaction. Estimations of paleobathymetry/paleoaltitude result from facies analysis and are indicated in comparison with the lake level noted 0. In this study, a paleobathymetry of 25 m is attributed to offshore environments, 10 m to transitional environments, 5 m to shoreface sediments, 0 to beach environments and a paleoaltitude of 10 m to fluvial plain and 20 m to alluvial fan. Here, a composite curve is reconstructed to represents the variation of accommodation between ~4.20 and ~1.28 Ma.

4. RESULTS

4.1. Facies associations

Five main facies associations (FA) are identified in the Nachukui Fm from ~4 to ~1.25 Ma revealing the existence of five groups of depositional environments (Table 1). Offshore (FA1) and transitional (FA2) deposits consist mostly of mudstone (i.e., pelites). In contrast to FA1, FA2 shows recurrent intercalated coarser grained beds, which we interpret as storm deposits,

indicating sedimentation in the transition zone (Coe et al., 2003) Wave-influenced nearshore (FA3) deposits are dominated by very well- to well-sorted sandstone, bioclastic sandstone to rudstone with bioclasts corresponding mainly to shells or shell fragments of aquatic mollusks. Within FA3, we recognize three sub-groups: the shoreface (FA3-a), the beach (FA3-b) and the backshore (FA3-c). River-dominated nearshore (FA4) deposits are mixed clastic and carbonate material deposited into coastal alluvial systems. River (FA5) deposits consist of mudstone, sandstone and conglomerate reflecting deposition attributed to the activity of river systems in the subaerial domain. Within FA5, two sub-environments are identified in FA5: the alluvial plain (F5-a) and alluvial fan (F5-b) sub-environments.

4.1.1. Offshore deposits (FA1)

Sediments of FA1 are made of blackish, brownish or greenish mudstone (Fig. 5A), depending on the locations. Prevailing sedimentary structures consist of mm- to cm-scale horizontal laminations (Fig. 5B), even if cm-scale low-angle ripple laminations are observed in places (Fig. 5B). Mudstones are commonly intercalated by dm- to m-thick massive diatomite beds (Fig. 5C). Occasionally, isolated dm- to m-thick tabular or lenticular massive or normal-graded sandstone beds characterized by erosive bases are embedded in mudstones (Fig. 5D). In places, such sandstone beds are sources for m-scale dyke-type donwtapering intrusions. Aquatic vertebrate remains such as fish bones and teeth are common in FA1.

Both mudstones and diatomite reflect suspension fallout from settling processes of fines in subaqueous low-energy depositional environments (Scholle, 1971; Boulesteix *et al*., 2019), even if occasional ripple laminations in mudstones suggest occasional occurrence of tractional currents. Predominance of settling processes reflects a depositional environment below both the fair-weather and the storm wave bases that define the offshore domain (Coe *et al*., 2003). Occasional evidences for tractional current in mudstones are attributed to

occasional bottom currents that can occur relatively frequently in offshore lake environments (e.g., Nutz *et al.*, 2014, 2015a, 2018). Isolated massive sandstone beds indicate debris-flow deposits (Talling *et al.*, 2012). Intercalated in offshore mudstones, these debris flows can be attributed to either river-derived underflows or mass-wasting turbidity currents that are frequent processes in deep portions of lakes (e.g., Zavala *et al.*, 2010; Nutz *et al.*, 2014). However, the occurrence of normal graded beds with only sparsely observed inverse graded intervals rather suggests mass-wasting gravity flows (Mulder *et al.*, 2003). Subsequently to their deposition, sandstone beds are occasionally parent units for injectites that may be conservatively attributed to seismicity (Hurst *et al.*, 2011). Previous estimates based on wave observations on modern Lake Turkana broadly support a storm wave base at a depth of ~ -20 m (Nutz *et al.*, 2017). Thus, FA1 is considered to reflect sedimentation in the lake, at >20 m of depth.

4.1.2. Transitional deposits (FA2)

Sediments of FA2 consist of cm- to dm-scale thick coarse-grained beds made of sandstone or shells that alternate with mudstone intervals similar to those observed in FA1. Two types of intercalated coarse-grained beds are identified. The first type of FA2 corresponds to dm-thick sandstones beds (Fig. 5E). Where internal primary sedimentary structures are obvious, their lower boundary is erosional, in places marked by aligned mm- to cm-scale gravels and pebbles. Overlying their lower boundary, coarse sandstones are horizontally laminated grading upward to medium sandstones in symmetric ripple laminations (Fig. 5E) showing a general fining-up interval. In places, the top surface show preserved cogenetic interferent ripples (Fig. 5F) organized in brick and tiles pattern (Allen, 1982). The second type of coarse-grained beds in FA2 consists of cm-scale thick erosionally-based beds (Fig. 5G) of shell

assemblages, either consisting of entire gastropod shells or fragments, forming massive packstone beds (Fig. 5H).

Similar to FA1, mudstone intervals separating coarse beds reflect subaqueous low-energy depositional environments characterized by predominant suspension fallout of fines from settling processes. Embedded sandstone beds marked by basal erosion associated with a fining-up trend indicate sedimentation from a high-density gravity flow (Talling *et al.*, 2012) and represent event beds (Seilacher, 1982). In the upper part, cogenetic interferent ripples evidence influence of interfering oscillatory currents during the waning of the flow (Aigner, 1985; Jelby *et al.*, 2019). In FA2, these sandstone beds are interpreted as storm beds (Johnson and Baldwin, 1996; Jelby *et al.*, 2019). The second type of coarser beds reveals *en masse* sedimentation (Talling *et al.*, 2012) of shells or shell fragments suggesting that repeated currents reworked shells in the nearshore domain to transport them to areas between the fair-weather and the storm wave bases. Thus, shelly coarse beds embedded in subaqueous mudstones are also attributed to storm-induced currents. Non-amalgamated tempestites usually form at depths between the fair weather and the storm wave bases. Considering a storm wave base around -20 m and a fair-weather wave base at around -5 m in modern Lake Turkana (Nutz *et al.*, 2017), the transitional depositional environment is here estimated within water depths of 5 to 20 m.

4.1.3. Wave-influenced nearshore deposits (FA3)

Shoreface deposits (FA3-a)

FA3-a deposits consist of well-sorted medium to coarse sandstone alternating in places with dm-scale packstone to rudstone beds composed of freshwater gastropod and bivalve shells in 1-10 m thick packages. Mollusk specimens are typically rolled and somewhat abraded, with the valves of bivalves regularly being separated. Planar laminations and wave ripple

laminations (Fig. 6A, B and C) prevail even if occasional dm-scale beds with planar cross-laminations are observed. In places, heavy minerals are abundant and concentrated in dark laminations (Fig. 6A). Occasionally, dm-scale sandstone beds laterally varying in thickness at m-scale distances reveal hummocky cross-stratifications (HCS) (Fig. 6D). Vertical burrows are common. When observed on large exposures, such sediments are included in 2 to 10 m-thick wedges that comprise large-scale oblique master beds gently dipping towards the basin (Nutz *et al.*, 2017).

Predominance of mollusks in bioclastic sandstone and packstone suggests nearshore environments. Rolled and somewhat abraded shells, with the valves of bivalves regularly being separated reveal a relative high-energy environment while wave-ripples reveal oscillatory currents. Associated with planar laminations, sedimentary structures reflect a deposition in the nearshore domain either in the surf or the swash zone (Clifton *et al.*, 1971; Massari and Parea, 1988; Clifton 2006). Occasional planar cross-laminations reveal dunes attributed to tractional currents in the build-up or surf zones (Clifton *et al*., 1971; Massari and Parea 1988) while HCS indicate the combination of large-scale oscillatory and tractional flows developed during storm events in the shoreface (Harms *et al.,* 1975; Dott and Bourgeois, 1982). Packstone and rudstone beds support a deposition in coastal domains. Finally, altogether these sedimentary structures evidence sedimentation in a wave-dominated shoreface depositional environment. Large-scale and low-angle master beds are interpreted as clinoforms revealing basinward progradations of coastal wedges (Nutz *et al.*, 2017). The shoreface domain is located between the fair-weather wave base and the foreshore and corresponds to that part of the shorezone where sediments are constantly agitated by waves. This interval is broadly estimated between -5 and -2 m in depth (Nutz *et al.*, 2017).

Foreshore deposits (FA3-b)

Depending of their location, FA3-b deposits consist of fine conglomerate, coarse sandstones, bioclastic sandstone, packstone or rudstone beds (Fig. 6E and F) composed of freshwater mollusk shells. Mollusk specimens are abraded and with the valves of bivalves are regularly separated. Two types of foreshore deposits are distinguished in measured sections. The first type of FA3-b is included in 1 to 2-m-thick sediment bodies made of planar laminations in low- (<5°) to moderate-angle (5-10°) oblique master beds dipping basinward (Fig. 6G, H). The second type of FA3-b consists of bioturbated coarse sandstone or bioclastic sandstone included in successive dm- to m-scale intervals. Each interval is characterized in the lower part by horizontal laminations with occasional planar and through cross-laminations with few burrows or rhizoliths. The abundance of such burrows or rhizoliths increases in the upper part which is pervasively bioturbated. Primary structures such as planar or cross-laminations, are disrupted and not obvious anymore. The upper part is topped by an irregular surface where burrows or rhizoliths are interrupted (Fig. 6I).

In the two types of FA-b, important accumulations of mollusk shells in bioclastic sandstone, packstone and rudstone evidence sedimentation in the coastal domain. In the first type, oblique master beds are interpreted as reflecting the accretion of successive beach faces (Dabrio, 1982; Tamura, 2012) developed in the swash zone (Clifton *et al.*, 1971) supported by the abraded shell assemblages that suggest the influence of waves. Low-angle master beds express relative high-energy dissipative beaches in a high-energy wave-dominated foreshore environment. Higher-angle oblique stratifications rather indicate more reflective beaches developed in a moderate-energy wave-dominated foreshore environment (Wright *et al.*, 1979). In the second type of FA3-b, the abundance of bioturbation reflects deposits from a low-energy shore environment where aquatic macrophytes were abundant and fairly impacted by waves. This type reflects sedimentation in a more sheltered zone, such as coastal swamps that experienced alternations between sedimentation and the development of vegetation. Both

versions of FA3-b represent a sedimentation in a foreshore depositional environment, which bathymetric range is broadly estimated between -2 and 0 m (Nutz *et al.*, 2017), corresponding to the swash zone in wave-influenced nearshore domains or to areas where bathymetry allowed aquatic macrophytes to develop.

Backshore deposits (FA3-c)

FA3-c deposits mostly consist of well-sorted medium sandstone in dm-scale beds alternating with bioclastic sandstone beds. These beds are included in 1–3 m thick sediment bodies (Fig. 7A), which show large-scale high-angle (>10°) oblique master beds dipping toward the basin margin, in places grading updip to sub-horizontal beds. Master beds are frequently reworked revealing recurrent reactivations (Fig. 7A). Commonly, the uppermost sub-horizontal beds display dm-scale burrows and rhizoliths (Nutz *et al.*, 2017).

Mollusk shells in bioclastic sandstone indicate nearshore deposits. Landward oblique master beds reveal migrating features from the central portion of the basin toward the margin whereas sub-horizontal beds in the uppermost part reflect coeval aggradation. In nearshore domains, m-scale landward-migrating features are interpreted as washover fans (e.g., Schwartz, 1982; Neal *et al.*, 2003; Baumann *et al.*, 2017; Nutz *et al.*, 2017). Internal reactivation surfaces and soil development at the top of sediment bodies indicate intermittent migration and distinct construction phases. During storm episodes, infragravity waves transport sediments from the foreshore to the backshore domain forming washover accumulations (Baumann *et al.*, 2017). Between two successive storms, soils developed on top surfaces (Nutz *et al.*, 2017). Washover fans are important features along wave-dominated coasts of lacustrine systems (Nutz *et al.*, 2018). In the Turkana Depression, early Quaternary, Holocene and modern examples of washover fans are abundant (Tiercelin *et al.*, 2010; Nutz *et*

al., 2017; Schuster and Nutz, 2018). They formed broadly between 1 and 5 m above the contemporaneous lake level.

4.1.4. River-dominated nearshore deposits (FA4)

FA4 deposits are represented by two different types. The first type of FA4 consists of poorly-sorted fine to medium matrix-supported conglomerates in dm- to m-thick beds. Matrix consists in sandstone, bioclastic sandstone and packstone including mollusk shells or shell fragments (Fig. 7B, and C). These beds are systematically intercalated in river deposits (*FA5*; Fig. 7C and D). Conglomerate clasts are relatively immature whereas the matrix commonly includes planar laminations and wave ripple laminations in sands, frequently associated with shells or shell fragments accumulations. Frequently thin concentric carbonate laminae coat conglomerate clasts to form cm- to dm-scale isolated oncolites (Fig. 7C and D). In places, adjacent oncolites form clast-supported conglomerates, topped by dm- to m-thick continuous beds made of upward growing domal stromatolites (Fig. 7E). The second type of FA4 deposits corresponds to intervals of several meters including dm- to m-scale very well-sorted sandstone beds that are either tabular of channelized with erosional lower boundaries (Fig. 7F, G) separated by dm-scale massive mudstone beds. Sandstones beds include planar laminations, occasional climbing-ripples and rare dm-scale cross-laminations.

In the first type, mollusks in bioclastic sandstone and packstone indicate nearshore environments while wave ripples and oncolites suggest the recurrent influence of waves. The poor sorting of conglomerates and sandstones and the low maturity of pebbles suggest the absence of coastal transport. As such, fluvial tributaries that supplied sediment in the lake were close to depositional areas. This type of FA4 therefore derives from gravity flows in the form of hyperconcentrated flows (Mulder and Alexander, 2001) and/or from debris flows (Talling *et al.*, 2012) that were subsequently partially reworked by waves forming fluvial-

induced deposits in the nearshore domain. During river-flood, rivers generated flows that directly entered the lake and deposited coarse sediments in the nearshore domain forming mouth bar complexes. Waves then partially reworked previously deposited material and mixed river-derived sediments with *in situ* biogenic carbonate nearshore deposits (i.e., mollusk shells and microbialites). This type of FA4 includes coastal alluvial fan deposits. In places, long-term abandonment or avulsion of river channels allowed washing and sorting of oncolites by waves and subsequent development of thick carbonate intervals made of domal stromatolites (Fig.7E). The development of the carbonate interval stopped when alluvial sedimentation restarted. The second type of FA4 deposits characterized by tabular or channelized sandstone beds reveals sedimentation in the form of unconfined or confined turbulent flows, respectively (Wright 1977). Stacking of several sandstone beds separated by dm- to m-scale mudstone intervals suggest recurrent flows that broadly affected broadly the same areas. This suggests the proximity of river mouths where river floods might have repeatedly generated hyperpycnal flows that supplied sands in the nearshore domain forming river mouth complexes. Finally both types of FA4 are interpreted as river-dominated nearshore deposits.

4.1.5. River deposits (FA5

Alluvial plain (FA5-a)

FA5-a deposits consist of tens of intervals of >10 m thick composed of sandy mudstone to muddy sandstone. These deposits include lenticular sediment bodies that are tens to hundreds of meters wide and meters thick, made of fine to medium conglomerates or coarse sandstones (Fig. 8A and B). In sandy mudstones to muddy sandstones intervals, primary depositional structures are difficult to observe, yet crude horizontal stratification seems to prevail. Pervasive termite nests are observed (Fig. 8C), in places connected to each other (Duringer *et*

al., 2007). Occasional ant galleries also occur (Tiercelin *et al*., 2010). Embedded lenticular sediment bodies are erosion-based, showing dm- to m-scale trough cross-laminations (Fig. 8D) broadly oriented basinward. Occasionally, matrix-supported fine conglomerates in dm-scale fining-up intervals are observed (Fig.8E). Bioturbation is frequent in the form of rhizoliths (Fig. 8F and G).

Sandy mudstones to muddy sandstones reflect deposition from both suspension fallout and tractional currents. The abundance of rhizoliths and social insect nests indicates terrestrial areas stable enough for recurrent soil development (plant growth and activity of soil organisms). Sandy mudstones to muddy sandstones reflect deposition of overbank fines during river overflows. During river overflow, sand is deposited on the alluvial plain. Subsequently, just after the peak of flood when the alluvial plain is flooded, mud deposited. Between two river flood events, soils developed that resulted in homogenization of previously deposited muds and sands to form the currently observed mixture of sandy mudstones to muddy sandstones. In coarse lenticular sediment bodies corresponding to channels infill, dm- to m-scale through cross-stratifications suggest migrating dunes and bars revealing tractional currents attributed to dilute flows during lower flow regime conditions (Allen, 1964, 1982). The occasional occurrence of fining-up intervals evidence occasional sedimentation from gravity flows while the presence of a muddy matrix suggests sedimentation from cohesive debris flows (Blair and McPherson, 1994, 2009). The large-scale lenticular geometry reveals that both dilute and debris flows were confined in wide and shallow channels. Finally, FA5-a deposits are interpreted to represent infill of braided river channels (Allen, 1964; Blair and McPherson, 1994); the broadly eastward directions of paleoflows suggest channels originating from the western rift shoulder. Thus, overbank deposits laterally associated with alluvial channels indicate organized stream flows that reflect alluvial plain systems (Miall, 1996; Bridge, 2003).

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Alluvial fan (FA5-b)

FA5-b are included in tens of meters thick and hundreds of meters wide sediment packages. Two types are distinguished. The first type of FA5-b is dominated by successive m-thick intervals of matrix-supported coarse conglomerate separated by dm- to m-scale beds of sandstones to fine conglomerates (Fig.8H). Coarse conglomerates show clasts ranging from pebbles up to boulders (Fig. 8I) while matrix corresponds to muddy coarse sands and gravels. Conglomerates are mostly massive even if faint horizontal laminations are occasionally observed. Intercalated, dm- to m-scale sandstones and fine conglomerates are moderately-sorted in horizontal stratifications. Common dm-scale long rhizoliths are observed, affecting all lithologies. The second type of FA5-b consists of dm-scale planar beds made of poorly-sorted coarse sandstones intervals sharply alternating with fine conglomerates intervals (Fig. 8J). Faint horizontal laminations prevail in all lithologies, in places associated with dm-scale trough-cross laminations. Locally, dm-scale intervals of medium conglomerates showing oriented and imbricated clasts are intercalated. Even if rhizoliths are also observed, their abundance is lower compared to that in the first type.

In the first type of FA5-b, massive matrix-supported conglomeratic intervals reflect debris flows (Blair and McPherson, 1994, 2009) whereas occasional intercalated sandstone beds indicate occasional deposition from dilute flows during upper flow regime conditions (Allen, 1964; 1982). Rhizoliths suggest repeated and durable interruptions of flows favoring vegetation development during non-depositional periods. The second type of FA5-b reveals a sedimentation dominated by tractional currents from dilute flows during upper flow regime conditions (Allen, 1964; 1982). In places, sedimentation that originated from dilute flows is associated with sedimentation from concentrated to hyperconcentrated flows (Mulder and Alexander, 2001). In both types, the rarity of channelized features and occurrence of sheet

form deposits evidence unconfined flows typical for sedimentation in alluvial fan systems. The first type is dominated by debris flows while the second type suggests sedimentation rather dominated by sheetflood processes (Blair and McPherson, 1994, 2009; Leleu *et al.*, 2005, 2009).

4.2. Sequence stratigraphy and high-amplitude paleolake fluctuations

Following the interpretation of depositional environments (FA1-FA5) and observation of their stacking patterns (Figs. 4, 9, 10, 11, 12), we delineated T-R sequences over the complete succession and propose an interpretative lithostratigraphic chart (Fig. 13). In total, 7 high-amplitude sequences (S1-S7) are delineated from ~1.00 to ~1.25 Ma. Their limits and compositions are presented hereafter.

Directly overlaying the Gombe Basalt, the succession starts either with alluvial fan deposits (*FA5*; sections 1, 2, 3 and 4) or offshore mudstones (*FA1*; section 5). In section 2, alluvial fan deposits are then overlain by 3-5 m thick transitional deposits whereas in sections 1, 3 and 4, they are directly overlain by 5-10 m thick offshore lacustrine deposits. Offshore lacustrine deposits grade upward to 1-4 m thick transitional (*FA2*) and/or shoreface (*FA3-a*) depositional environments (sections 3 and 4). This succession reveals a lake transgression onto subaerial alluvial fans or directly onto the Gombe Basalts (Fig. 9A). Above, the transitional (*FA2*) and shoreface (*FA3*-b) deposits indicate a subsequent shallowing trend attributed to a lake regression. The top of shoreface deposits in sections 3 and 4 marks the upper boundary of the sequence S1 (Fig. 4) and thus the lower boundary of S2 (SB2). The maximum flooding interval of sequence 1 (MFI1) is estimated in the offshore lacustrine deposits.

Overlying SB2, 10-15 m thick offshore lacustrine deposits *(FA1)* are observed in sections 3 and 4. Upward, they grade to either 1-3 m thick river-dominated nearshore (*FA4*;

section 3), shoreface (*FA3-a*, section 4) or transitional deposits (*FA2*) characterized by shelly storm beds (section 5) and then, in sections 3, 4 and 5, to 10-15 m thick river deposits (FA5) (Figs. 4, 9B and C, 10A and B). Section 3, 4 and 5 show a rapid deepening of the lake above SB2 followed by the progressive shallowing and an emersion represented by river deposits. In sections 4, river deposits (*FA5*) are directly erosionally overlying a river mouth complex (F4) and offshore deposits (*FA1*), infilling incised valleys (Fig. 10D and E). In section 5, river deposits show faint clinoforms that dip toward the central portion of the basin (Fig. 10A) revealing a progradational trend in this package. In sections 3, 5 and 6, river sediments are topped by an erosional surface underlined by aligned pebbles, frequently coated by microbialites (Fig. 10C). Above this erosional surface, dm- to m-scale beach deposits (*FA3-a*) are observed in section 3 (Fig. 11A), overlain by 20-30 m thick offshore deposits (*FA1*) marking the onset of S3 (Fig. 11B). In sections 5 and 6, offshore deposits directly overlay river deposits (*FA5*). In all sections, the upper boundary of river deposits (*FA5*) constitutes the top of S2 and, therewith, the sequence boundary of sequence S3 (SB3). Located in more distal part, section 7 is mostly composed of offshore material (*FA1*) occasionally intercalated by deposits from either transitional (*FA2*) or nearshore (*FA3*) depositional environments. In section 7, two sequences and the onset of a third one are also delineated supporting interpretations in more proximal domains.

Indeed, rirectly above SB3, ~30-40 m of offshore mudstones (*FA1*) are observed (sections 6, 7 and 8; Fig. 11B). Upwards, offshore mudstones grade to m-scale sandstones and conglomerates deposited in river-dominated nearshore (*FA4*) environment, in places associated with storm deposits. Overlaying, river-dominated nearshore deposits progressively grade to a 90 m thick interval of river deposits (*FA5-a*), which show alternations of overbank deposits, alluvial channels that grade upward to alluvial fan deposits (Fig. 11C). For the same fluvial interval, Harris et al. (1988) report 160 m, before reaching the Lomekwi Tuff.

However, even if we did not observe such thickness on field, description by these authors reveals broadly similar river deposits all along the succession and as such, do not imply changes in our sequence delineation. Indeed, the progressive transition from offshore (*FA1*), to transitional (*FA2*), river-dominated nearshore (*FA4*) and then river (*FA5*) deposits reveal a general progradation. This sequence constitutes the regressive upper part of S3 (Fig. 4) and is topped by SB4. The MFI3 is estimated slightly below the Tulu Bor Tuff.

Alluvial plain deposits with a thickness of 35-40 m (section 9) directly overlie SB4. These sediments represent predominantly overbank deposits in alluvial plain environment (*FA5-a*; Fig. 11D) and they include the Lokalalei and Kokiselei Tuffs. In the upper part of section 9, two intervals of nearshore deposits (*FA3-b*) intersperse these river deposits (Fig. 11E). The second nearshore interval is supposed to coincide with the 7-8 m thick basal nearshore interval observed in the lowermost part of section 10 (Fig. 4). In this section, nearshore deposits correspond to low-energy coastal swamp (*FA3-b*). Upwards, a 25 m thick package of alluvial plain deposits is overlain by a m-scale nearshore (*FA3*) interval grading upward to shoreface (*FA3-a*) and then offshore mudstones (*FA1*). Offshore mudstones (*FA1*) are 40-45 m thick (Fig. 11F), in places intercalated by shoreface deposits (*FA3-a*). This succession forms the transgressive part of S4.

In section 12, about 20 m-thick wave-dominated nearshore deposits (*F3*) are identified (Fig. 12A) above the offshore mudstones, evidencing a contraction of the lake (Nutz *et al.*, 2017). This contraction started ~15 m below, where the maximum flooding interval of S4 is estimated, and corresponds to the regressive part of S4.

Upward, nearshore deposits are overlain by transitional deposits (*FA2*) grading to offshore mudstones (*FA1*) marking the onset of a subsequent transgression attributed to S5 (Fig. 4). In section 13, the regressive part of S4 is not observed. Here, above the KBS tuff, 65 m thick of offshore mudstones (*FA1*) are present. They grade upward to 20 m thick

transitional deposits (*FA2*) into which the Etir Tuff is identified. In the uppermost part of section 13, 15 m-thick river-dominated nearshore (*FA4*) and shoreface (*FA3-a*) sediments are observed, which are correlated with the lowermost part of section 14. This succession records a shallowing attributed to the regressive part of S5.

Above SB6, 40 m of offshore mudstones (*FA1*), reveal a subsequent transgression attributed to the lower part of S6 (Fig.4). Hence, the top of shoreface deposits (*FA3-a*) at the bottom of section 14 is interpreted as the top of S5 and the lower limit of S6 (SB6), this long-term regressive trend of S5 marks a contraction of the lake before another transgression that represents an expansion of the lake. Upward, offshore deposits are overlain by a 50-60 m thick interval of river-deposits, commonly embedded by shoreface (*FA3-a*) and river-dominated nearshore (*FA4*) intervals (Fig. 4). Overlying, 20 m thick offshore mudstones (*FA1*) evidence a subsequent transgression (Fig. 12B, C). The interval of river deposits reflects an emersion before another subsequent flooding. As such, this interval marks the regression associated with the upper part of S6 and the transgression coeval with the lower part of S7 (i.e., SB7).

Sections 17 and 18, which are located closer to the MRFL fault indicate that the offshore mudstones are capped by a 120 m thick interval that consists of alluvial plain deposits (FA5-b; Fig. 12D, E and F), in places intercalated by m-scale river-dominated nearshore deposits (Fig. 12D, E). This interval corresponds to the regressive part of S7. Although our sections end ~15 m above the Middle Nariokotome Tuff, at ~1.25 Ma, data in Harris suggest an interval of river-derived deposits similar to FA5 embedded by common microbial constructions (oncolites) reflecting river-dominated nearshore deposits (FA4) until ~0.75 Ma. According to Harris et al., (1998), progradation of such alluvial system remained limited with a shoreline that has fluctuated between 2 and 4 km far from the MRLF.

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The interval between ~4 and ~3.44 Ma includes the Lonyumun and Kataboi Members. This interval records two high-amplitude sequences (S1 and S2) and the lower part of a third high-amplitude sequence (S3) revealing two major T-R cycles and a subsequent Transgression (Fig. 13). The shoreline fluctuated from the MRLF to ~8 km eastward. Lying directly onto the Gombe Basalts, the lowermost part of S1 is estimated at ~4 Ma. This time frame is further supported by the presence of the Moiti Tuff (3.97 Ma) in offshore mudstones (section 7), revealing that lacustrine conditions existed prior to this age. The top of S1 is not dated and no tuff is available to provide an absolute age. However, the regressive part of S2 is estimated around 3.6 Ma as revealed by the K82-742 and Loruth Tuffs identified in section 3 (Table 2). Above, the transgression associated with S3 reached its maximum slightly after 3.44 Ma (Fig. 4). In section 7, the presence of deposits *(FA1, FA2 and FA3-b)* that formed under water during the entire considered time interval shows that a lake continuously occupied the basin during this period even if fluctuations occurred. From 3.44 Ma to 1.89 Ma, the Lomekwi, Lokalalei and Kalochoro Members record the upper part of sequence S3 and the lower part of sequence S4 revealing a long-term regression followed by a subsequent transgression. Transition from the regressive trend of S3 to the transgressive trend of S4 is estimated to be slightly older than the Lokalalei Tuff (i.e., slightly older than ~2.53 Ma) where alluvial plain deposits become predominant. As such, the long-term regressive trend associated with S3 is estimated between ~3.44 and ~2.53 Ma while the long-term transgressive trend of S4 is estimated to start after ~2.53 Ma. The maximum flooding interval of S4 (i.e., MFI4) is estimated ~15 m below the KBS Tuff, marking the onset of the regressive part of S4 slightly before 1.89 Ma. Close to the border fault, 10 m-thick wave-dominated deposits (F3) embedded in river deposits reflect this maximum flooding interval (Fig. 11G). Superimposed, three higher-frequency sequences are observed during the long-term transgression of S4 (Figs. 4, 13). From ~1.87 to ~0.75 Ma, the Kaitio, Natoo and Nariokotome Members record the

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upper part of sequence S4 and 4 sequences referred to as S5, S6 and S7 (Fig. 4). Limits of sequences are particularly uncertain in this interval, however SB5 is estimated at ~1.76 (Nutz *et al.* 2017) while SB6 and SB7 are estimated slightly younger than ~1.44 Ma and between ~1.43 and ~1.4 Ma, respectively. The maximum flooding interval (i.e., MFI7) of S7 is estimated at ~1.4 Ma while the regressive part of S7 likely includes all the succession until 0.75 Ma, even if subordinate-order cycles might have occurred.

4.3. Sedimentary systems

Combining the interpretation of depositional environments and their lateral relationships (Fig. 13), two sedimentary systems reconstructed in the West Turkana area are presented and referred to as type-1 and type-2 (Fig. 14). Each sedimentary systems represents the coexistence of different depositional environments along an ideal downslope transect from the border fault to the central portion of the basin. These sedimentary systems alternatively developed between 4.00 and 1.25 Ma as sedimentation varied through time in the West Turkana area.

The type-1 sedimentary system is characterized by the predominance of wave-influenced coastal deposits. From the border fault to the central portion of the basin, the type-1 sedimentary system consists of a few km-wide river deposits near the border fault. Where observed, river deposits form alluvial fans mostly made of fine conglomerate to gravelly sandstones indicating sheetflood-dominated alluvial fans (*FA5-b*), which are laterally associated with alluvial plain deposits (*FA5-a*) showing alternations of overbank deposits and channels. Alluvial fans are fed by rivers draining the western rift shoulder. Basinward, river systems grade rapidly to thick coastal wedges that consist of wave-dominated nearshore deposits (*FA3*) such as shoreface, beaches and washover fans that form wide strandplain.

Finally, wave-dominated nearshore deposits connect laterally to storm-dominated transitional (*FA2*) and then to offshore deposits (*FA1*). The type-1 sedimentary system is characterized by (i) the limited development of alluvial fans along the border fault and (ii) the important development of wave-dominated coastal wedges. Type-1 sedimentary systems are conspicuously recognized between ~4 and ~3.40 Ma and between ~2.5 and ~1.42 Ma (Fig. 13).

In contrast, the type-2 sedimentary system is characterized by the predominance of river deposits. From the border fault to the central portion of the basin, type-2 is characterized by wide conglomeratic alluvial fans which have been deposited mostly by debris flows. Laterally, alluvial fans grade to an alluvial plain (*FA5-a*) and then to river-dominated nearshore deposits (*FA4*). Toward the basin, these deposits rapidly grade to offshore mudstones (*FA1*). Finally, in opposition to type-1, type-2 sedimentary system are characterized by the development of (i) wide and thick river deposits along the border fault in the form of alluvial fans and plains, (ii) river-dominated nearshore deposits in more distal areas and (iii) a scarcity of wave dominated coastal wedges and features. The type-2 sedimentary system is observed during periods between ~3.40 and ~2.5 Ma and after ~1.42 Ma (Fig. 13) in thick packages associated with regressive trends during periods of long-term progradation.

Fundamental differences between type-1 and -2 sedimentary systems derive from the competition between the development of wave-dominated coastal and river-dominated systems in West Turkana. Fluctuation in the amount of sediments entering the basin from the western rift shoulder likely controlled the shifts from one system to the other. Indeed, during periods of reduced sediment supply, alluvial systems were smaller and processes related to waves and alongshore drift were dominant forming successive prograding and retrograding strandplain systems at the origin of type-1 sedimentary system. Lake level fluctuations were

the main mechanism to drive transgression-regression cycles revealing dynamics of an "accommodation-dominated system" (Neal and Abreu, 2009; Zhang *et al.*, 2019). In contrast, during periods characterized by an important sediment supply, alluvial systems developed and prograded toward the basin forming quasi-continuous regressive intervals regardless paleolake fluctuations. The flux of sediments towards the basin inhibited the expression of wave-related processes, except perhaps during infrequent short and abrupt lake level rises. Hence, type-2 sedimentary system rather reveals dynamics associated with "supply-dominated systems" (Burgess and Hovius, 1998; Carvajal and Steel, 2006; Zhang *et al.*, 2019).

4.4. Paleovegetation changes

Both the Nachukui and the Shungura records of $\delta^{13}C_{pedogenic\ carbonate}$ exhibit very widespread values (Fig. 15). Ranging from -10‰ to 0‰ these values indicate heterogeneity in the vegetation, i.e. closed and open habitats likely co-occurred within a given area, as is observed today for the riparian forest along the Omo River in an otherwise open savanna-dominated landscape. After data processing, marked trends emerge from the $\delta^{13}C$ records (Fig. 15). The isotopic record of soil carbonate from the Nachukui Fm shows an almost steady increase in C_4 biomass, suggesting gradual aridification from ~3.90 Ma to ~0.90 Ma in West Turkana area. A significant biome change occurred at ~3.25 Ma, when woodland/bushland/shrubland were replaced by wooded grassland, suggesting a change in water supply as woody cover dropped below 40%. The woody cover at West Turkana reached its smallest extent at ~2.00 Ma, and between ~1.50 and ~1.30 Ma. In the Omo Valley, $\delta^{13}C$ record of the the Shungura shows a different pattern. Interestingly, the Omo Valley exhibits a much longer record of woodland/shrubland/bushland than West Turkana, which lasted until ~1.85 Ma. Woody cover was always greater in the Omo Valley than in West Turkana until ca 1.85 Ma, suggesting

greater water supply through rainfall, overbank floods or as groundwater in the Omo River alluvial plain. After 1.85 Ma, both records indicate wooded grassland with inferred woody cover not exceeding 20-40%. Our results thus show that like West Turkana area, the Omo Valley underwent a long-term aridification between ~3.20 and ~1.25 Ma. However, changes in the woody cover as inferred from changes in the $\delta^{13}C$ records seem to have been more gradual in West Turkana than in the Omo Valley. In the Omo Valley, periods with maximum woody cover of about 50/60 % occurred twice, in the lowermost part of the record at ~3.20 Ma and between ~2.25-2.00 Ma revealing particularly wet conditions. Although less pronounced than later at 1.65 Ma, somewhat drier conditions seem to have reduced tree cover between ~2.70 and ~2.50 Ma.

4.5. Forcing factors

4.5.1. Origin of lake-level fluctuations

Lake level fluctuations can be forced at various timescales by many factors, including climate, tectonism or volcanism (Tiercelin, 1990; Cohen, 2003). Changes in the precipitation/evaporation ratio (i.e., P/E ratio) can drastically modify lake level, especially in systems with high climate sensitivity such as the East African Rift System (Street-Perrott and Harrison, 1985; Trauth et al., 2010). Modifications of river catchments by tectonic events (i.e., block rotations, development of faulted barriers) can also modify the water balance of a lake. Furthermore, modifications of the basin physiography in active rift systems can also significantly modify lake levels at constant P/E ratio. Such modifications may occur by variation in subsidence through time, either tectonically or by repeated subsidence pulses related to volcanism. Indeed, important volcanic events potentially lead to either adjustment of the crust after successive emptying of a magma chamber under the basin or isostatic

adjustment caused by magma infill of significant proportions of the basin. Moreover, tectonism and volcanism can change significantly the position and threshold elevation levels of a lake outlet. In case of an overfilled lake configuration (*sensu* Carroll and Bohacs, 1999), the lake level is modified. Given the potentially intricate interactions among climate, tectonism and volcanism on lake levels through time, the respective contributions of these factors need to be reconstructed through time to understand the causes of high-amplitude fluctuations.

In the West Turkana area, evolution of the woody cover (Nachukui $\delta^{13}C$ record) does not match the reconstructed evolution of the paleolake water-level (Fig. 15). Major lake highstands are not coeval with any remarkable negative shifts in $\delta^{13}C$ values nor any particular periods of greater woody cover. It is therefore unlikely that rivers activity in the West Turkana area and local paleohydrological conditions onto the western rift shoulder had any significant control on the paleolake fluctuations. In opposition, evolution of the woody cover in the Omo Valley (Shungura $\delta^{13}C$ record) shows that major lake highstands between 3.2 and 1.25 Ma, namely part of the Lokochot, Lorenyang and Nachukui highstands established during periods of particularly high woody cover (negative shifts of $\delta^{13}C$ values), while the major Lomewki lake lowstand occurred when woody cover was low (positive shift of $\delta^{13}C$ values). Moreover, the interval between 1.9 and 1.5 Ma characterized by a general reduced extent of the paleolake, even water-level fluctuations occurred, is coeval with an interval of particularly lower woody cover (positive shit of $\delta^{13}C$ values). Such a match between the reconstructed paleolake water level and woody cover in the Omo River floodplain strongly supports that the activity of the Omo River controlled major fluctuations of the paleolake. Indeed, greater woody cover in the Omo Valley can only be achieved with enhanced regional rainfall over the river catchment and the resulting augmentation of groundwater in bigger, larger, and shallower aquifers. We conclude that rainfalls over the

Ethiopian Dome controlled high amplitude paleolake fluctuations at least between 3.2 and 1.25 Ma, and by extension all along the considered time interval due to a similar source configuration before and after that period. This is supported by the modern and recent configurations, as the Omo River provides 80-90% of water entering the modern Lake Turkana (Yuretich and Cerling, 1983; van der Lubbe *et al.*, 2017), with a lower proportion that nevertheless exceeded 50% (van der Lubbe *et al.*, 2017) during the African Humid Period. Finally, by confronting reconstructed evolution of paleolake level and evolution of woody cover, we demonstrate that climate was the dominant factor that controlled the high-amplitude paleolake fluctuations. Considering that superimposed higher-frequency paleolake fluctuations were also controlled by climate (Joordens *et al.*, 2011; Nutz and Schuster, 2016; Nutz *et al.*, 2017), climate was the preponderant factor at the origin of paleolake fluctuations during the Plio-Pleistocene.

4.5.2. Origin of type-1 and type-2 sedimentary systems in West Turkana

The repeated alternations of type-1 and type-2 sedimentary systems (Fig. 14) in West Turkana mostly derived from changes in the amount of sediment supplied to the basin from the western rift shoulder. Increased sediment supply from the western rift shoulder may have two potential origins: (i) climate, as increased rainfall increases the transport capacity of rivers, and thus influences the amount of sediment entering the basin or (ii) tectonic, as a pulse of the rift shoulder uplift and rejuvenation of the relief can enhance denudation and increase both size and amount of sediments transported by the rivers into the basin. The Nachukui $\delta^{13}C$ record shows that during the two intervals of increasing sediment supply that led to the development of the type-2 sedimentary system (i.e., ~3.40-2.50 Ma and after ~1.42 Ma), woody cover in West Turkana decreased (Fig. 15). Hence, the increased sediment supply observed in West Turkana between ~3.40-2.50 Ma and after ~1.42 Ma cannot be attributed to

increased rainfall. Based on that, we attribute the most part of increased sediment supply to pulses of the rift shoulder uplift and the resulting increase of denudation. For the period between ~3.40-2.50 Ma, which is coeval with an aridifiation, we cannot rule out a minor influence of the increased erosion deriving from the likely reduced vegetation onto the rift shoulder. However, this process is not invoked for increased sediment supply after 1.42 Ma, which is coeval with wetter conditions compared to the former configuration.

As a conclusion, type-1 sedimentary system reflects a sedimentary system controlled by lacustrine processes responsible for the redistribution of clastics associated with low to moderate sediment supply delivered by transverse rivers coming from a moderately-elevated rift shoulder. Paleolake fluctuations are expressed by successive transgressions and regressions controlled by rainfall fluctuations over the Ethiopian Dome and the catchment of the Omo River. The modern western margin of the North Lake Basin in the Turkana Depression, which displays a remarkable serie of strandplains, spits and wave-dominated delta evidencing massive reworking of clastics by coastal waves and alongshore drift (Nutz and Schuster, 2016; Schuster and Nutz, 2018), is a relevant analogue of type-1 sedimentary system (Fig. 14). In contrast, type-2 sedimentary system was mostly controlled by a pulsed uplift of the rift shoulder. Erosion of this newly-formed relief provided important sediment supply that led to a continuous progradation even the probable fluctuations of the lake, likely except during particular short-term significant lake level rises. The modern western margin of the Chew Bahir basin (southern Ethiopia) bordered by well-developed alluvial fans fed by sediments coming from the Hamar Ranges is an accurate modern analogue of type-2 sedimentary systems (Fig. 14).

5. INTERPRETATION AND DISCUSSION

5.1. Paleoenvironmental reconstructions - synthesis

Using sedimentological sections, panorama interpretations, and compiled $\delta^{13}C$ in soil carbonate datasets, we investigated paleoenvionmental changes in the West Turkana area. Seven key periods are represented reflecting different paleolandscapes (Fig. 16). Each period is characterized by specific sedimentary dynamics, lake level and woody cover reflecting various combinations of water and sediment supply from both the Western Turkana and the Omo Valley areas.

From ~4.00 to ~3.60 M, a paleolake highstand referred to as the Lonyumun Highstand established in the northern Turkana Depression (Fig. 16A). In West Turkana, the lake shore reached the fault scarp and, at least locally, water invaded downstream parts of the valleys of the rift shoulder. Water was fresh (Cerling, 1979; Van Bocxlaer, 2020). Sedimentation was dominated by wave-related coastal processes (type-1 sedimentary system) mixing clastic and carbonate (mollusks) material. A low to moderate amount of sediment was supplied from the moderately-elevated rift shoulder and the system was accommodation-dominated. Alluvial systems along the border fault were poorly developed, and it is likely that little space was available between the lakeshore and the fault scarp to allow an extensive strandplain to develop even under prevailing wave-dominated processes. The woody cover is estimated around 55-60% indicating significant activity of rivers coming from the rift shoulder and thus that rainfall was important over the West Turkana relief.

Between ~3.60-3.50 Ma, the lake level abruptly dropped causing a lake lowstand, i.e. the Lokochot Lowstand (Fig. 16B). Following the shoreline retreat, rivers entered the basin and reworked previously deposited sediments forming incised valleys. During this period, the area was non-depositional. This pronounced lake regression occurred while the woody cover in West Turkana was high, around 50%, suggesting that rainfall in this area and/or over the Lappur Range at that time did not decrease majorly. It is possible that the short-term lake

level drop reveals an abrupt reduction of rainfall in the catchment of the Omo River, but as no data on woody cover from the Omo Valley around ~3.50 Ma are available, the issue cannot be verified. After ~3.50 Ma, a lacustrine transgression occurred that reflects the progressive establishment of the Lokochot Highstand, which reached its largest extent slightly before ~3.44 Ma. The water was fresh to slightly brackish (Cerling, 1979). Yet, with a woody cover of 45%, this period was still particularly humid in West Turkana.

From ~3.40 to ~2.60 Ma, the paleolake experienced a progressive regression associated with a change of sedimentation mode (Fig. 16C and D). The sedimentation became primarily controlled by river deposits and a supply-dominated system developed resulting from the increased elevation of the rift shoulder favoring the intensification of erosion and thus of the sediment production. As such, the sediment supply from rivers that drain the rift shoulder drastically increased. A belt of alluvial systems developed along the border fault (Fig. 16C), and progressively prograded toward the central portion of the basin pushing the shoreline basinward regardless lake-level fluctuations, likely except during occasional short-term and particularly important phases of lake level rises. Coeval to the progressive fall of the lake level, woody cover decreased in West Turkana (from 45 to 35 % between 3.40 and 2.60 Ma) and in the Omo River catchment (from 65 to 45 % between 3.20 and 2.60 Ma) revealing significant rainfall decrease in both areas. This period is also characterized by an important diminution of δA favoring faster progradation. Altogether, alluvial belt progradation, lake level fall and reduced rate at which ccommodation space is created led to a drastic retreat of the shoreline, likely as important as during the Lokochot Lowstand. The paleolake reached its lowest extent at around ~2.60 Ma, which marks the peak of the Tulu Bor Lowstand. At that time, a >8 kilometers wide river-dominated system (Fig. 16D) characterized the Western Turkana area.

From ~2.60 to ~2.10 Ma, despite minor water-level fluctuations, the paleolake progressively extended. The mode of sedimentation changed with a progressive shift from river deposits (type-1 sedimentary system) to wave-dominated deposits (type-2 sedimentary system). This change suggests that the pulse of rift shoulder uplift became interrupted and the intensity of relief erosion progressively became less intense. At the same time, woody cover in the Omo Valley increased largely (from 45% to 65%), although it continued to decrease in the West Turkana area (from 35% to 25%). Hence, whereas conditions in the catchment of the Omo River humidified, the ongoing aridification continued in West Turkana. Between ~2.10 and ~1.90 Ma, the paleolake reached a maximum extent referred to as the Lorenyang Highstand (Fig. 16E). Sedimentation is wave-dominated with mixed clastic and carbonate (mollusk shells) material. A limited amount of sediment was supplied from the moderately-elevated rift shoulder and alluvial systems along the border fault were poorly developed.

From ~1.90 to ~1.85 Ma, the paleolake shrank until it reached a lake lowstand between ~1.85 and ~1.75 Ma (Fig. 16F). Sedimentation was almost exclusively wave-controlled, with carbonate-dominated materials (mollusk shells) and more limited clastics. Water became more brackish (Cerling, 1979). Alluvial systems along the border fault were very poorly developed as a very low amount of sediment was supplied from the moderately-elevated rift shoulder. This regression was accompanied by a drastic decrease of the woody cover in the Omo Valley (from 60 to 35%). In the West Turkana area, however, woody cover increased slightly (from 25% to 30 %) (Fig. 15). This marks a transition to drier conditions onto the Ethiopian Dome and slightly wetter conditions in the West Turkana area. From 1.75 to 1.45 Ma, the paleolake experienced two limited expansions separated by two contractions, while the woody cover in both the West Turkana and the Omo Valley areas ranges between 30 and 35%. Hence, although some fluctuations occurred, reduced rainfall over the Ethiopian Dome substantially constrained the extent of the paleolake.

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Around ~1.40 Ma, the paleolake reached another significant lake highstand referred to as the Nachukui Highstand. Water was brackish (Cerling, 1979). This highstand is coeval with greater woody cover and, hence, rainfall in the Omo River catchment area, but drier conditions in West Turkana. Slightly before 1.40 Ma, the mode of sedimentation changed and became river-dominated with abundant clastics associated with recurrent microbial oncolites and stromatolites. An alluvial belt developed along the border fault and progressively prograded toward the central portion of the basin. Comparably to the period between ~3.40 and ~2.50 Ma, at ~1.40 Ma, a new pulse of the rift shoulder uplift led to increased erosion and a supply-dominated river system was established (Fig. 10C). However, as opposed to the ~3.40–2.50 Ma period, the increasing δA (up to 425 m Ma^{-1}) around ~1.40 Ma probably inhibited extensive lateral progradation and major shoreline migration into the basin. Rather, it favored aggradation, which is confirmed by the coastal river deposits embedded in river deposits observed close to the border fault around ~1.25 Ma, showing limited shoreline migration between ~1.40 and ~1.25 Ma. According to Nariokotome Member architectures reported by Harris *et al.*, (1988), this long-term progradation appears to have continued until ~0.75 Ma, even if occasional short-term transgression occurred.

5.2. Paleoclimate implications

Precipitations over the Turkana Depression and the Ethiopian Dome are controlled by the annual latitudinal migration of the Intertropical Convergence Zone (ITCZ) between 15° north and 15° south, seasonally delivering moisture from both the Indian Ocean and the Atlantic Ocean (Levin *et al.*, 2009; Nicholson, 2017). Precipitations, however, drastically differ in the Turkana Depression and the Ethiopian Dome, and this difference is attributed to local factors (Nicholson, 2017) such as (i) the relief, south-west of the Turkana Depression, which is supposed to limit the flow of moist air from the Congo Basin and to create lee rain shadows

(Nicholson, 2016), and (ii) the recurrent Turkana jet, which significantly modulates precipitation in the Turkana Depression (Nicholson, 2016, 2017). Hence, whereas precipitation in the Turkana Depression reflects local patterns, precipitation over the Ethiopian Dome may directly respond to larger-scale and global climate processes. Our study shows that changes in the water-level of the paleolake are closely associated with changes in the Omo Valley woody cover, which are indirectly related to precipitation changes over the Ethiopian Highlands. Thus, the paleolake appears to have responded to regional and global climate processes and thus represents a valuable archive for the Plio-Pleistocene paleoclimate in the horn of Africa.

Numerous studies demonstrate that variations of insolation constituted major controls on fluctuations of lake levels during the Plio-Pleistocene in East Africa. By modifying intensity of rainfall, precession cycles have controlled well-preserved lake fluctuations during the Plio-Pleistocene period (Deino *et al.*, 2006; Ashley, 2007; Nutz *et al.*, 2017) and the Holocene (Larrasoana *et al.*, 2013; Shanahan *et al.*, 2015). Increase of insolation due to changes in precession parameters periodically led to increased intensity of monsoonal precipitation. Other studies (Trauth *et al.*, 2005, 2007) suggest that evolution of the eccentricity also is an important factor that impacted climate by modulating precession cycles. During periods of eccentricity minima, the amplitude of insolation peaks is lower and, as a result, the intensity and migration of the summer monsoon front is supposed to be drastically reduced over periods of 100 to 400 kyr. Thus, during periods of 100 to 400 kyr low eccentricity, impact of precession is reduced and does not allow significant humid periods to establish. Precession cycles generated paleolake level fluctuations of a few tens of meters during the Plio-Pleistocene in the Turkana Depression (Nutz *et al.*, 2017). However, the systematic impact of eccentricity is still debated. Boës *et al*., (2018) propose that eccentricity is a potential parameter at the origin of paleloake fluctuations, but this study concerns a

limited time interval only. By confronting reconstructed fluctuations of the paleolake and the evolution of eccentricity between 4.00 and 0.75 Ma (Fig. 15), only a partial correlation appears to exist between long-term lake highstand or lowstand and particular periods of 100 to 400 kyr eccentricity maxima and minima. Indeed, correlations between periods of low-eccentricity and lake lowstands or high-eccentricity and lake highstands are not systematically observed. The humid period between 4.00 and 3.60 Ma is characterized by high eccentricity, which could suggest a direct control by insolation. However, between 3.20 and 2.90 Ma, 2.40 and 2.10 Ma or 1.90 and 1.70 Ma, similar high eccentricity patterns do not coincide with the establishment of lake highstands. Conversely, although the low eccentricity period between 2.90 and 2.40 Ma is coeval with dry conditions, periods between 3.40 and 3.20 Ma or 2.10 and 1.90 Ma show relative low eccentricity and relative high lake levels. A direct influence of eccentricity on precipitation is thus not systematically observed and it is likely that direct insolation mostly modulated the climate of pre-established wet and dry periods derived from particular global climate events or transitions.

Between ~4.00 and ~3.40 Ma, a global humid period supposedly affected East Africa. This humid period is associated with the latest time of the "lower Pliocene forest expansion" period described by Bonnefille (2010), which constituted a global humid period in Africa between ~6 and ~3.6 Ma with a climax centred at ~4 Ma. According to Bonnefille (2010), this climax of humidity may have originated from the closure of the Isthmus of Panama between 4.70 and 4.20 Ma (Haug, 1998) that led to major changes in Pacific and Atlantic waters organization and to a southwards shift of the ITCZ (Billups *et al.*, 1999). However, more recent literature questions this explanation (Montes *et al.*, 2015) proposing that the Isthmus of Panama may have closed much earlier, about 15 Myr ago. Following the humid period, the regression of the paleolake between 3.40 and 2.60 Ma coincides with a progressive transition to a dry and cool period onto the Ethiopian Dome, estimated between 2.70 and 2.50 Ma by

Bonnefille (1983) and supported at continental-scale by reconstructions of Grant *et al.* (2017) (Fig. 15). Two origins are proposed for this dry and cool period, the first is the direct impact in tropics of eccentricity-driven insolation changes following the mechanisms mentioned above (Trauth *et al.*, 2009); the second is the onset of the Northern Hemisphere Glaciation (deMenocal, 2004), which is coeval with aridification in some places in the tropics (Bonnefille, 1983; Bobe *et al.*, 2002). Although we cannot resolve this issue, we suggest that according to the intensity of this phase of aridification, a combined effect of both factors could be envisioned. Subsequently, the Lorenyang Highstand reveals newly established wetter conditions onto the Ethiopian Dome with a climax from ~2.10 to ~1.90 Ma. The origin of this newly identified humid period on the Ethiopian Dome is unknown although coeval low-eccentricity argues against an influence of direct insolation-driven increases of precipitation. The end of this humid period and the transition to drier conditions onto the Ethiopian Dome after ~1.90 Ma coincides with the establishment of a stronger Pacific Walker circulation (Ravelo *et al*., 2004) which has been proposed to have led to drier climate in Africa between ~1.90 to ~1.50 Ma (Trauth *et al.*, 2009). Finally, at around 1.40 Ma, a new important humid period reflects renewed wetter conditions. Although the global origin of such humid period is not identified, it coincides with a period of relatively high eccentricity which could have favoured a regional humidification. By the end, climate processes that influenced precipitations onto the Ethiopian Highlands, and thus fluctuations of the paleolake in the Turkana Depression, derive from the combination of global climate events modulated by the direct impact of insolation through the interaction of precession and eccentricity cycles.

5.3. Did a single long-lived lake occupy the Turkana Depression between 4 and 0.75 Ma?

Early paleoenvironmental studies (Cerling, 1979; Williamson 1981) proposed that the Turkana Depression has been occupied from ~4.00 to ~0.75 Ma by a single long-lived lake,

but this paleoenvironmental reconstruction was later adjusted(Brown and Feibel, 1988, 1991). The latter authors suggested that sedimentation during this period was for 85% of the time river-dominated, occasionally interspersed by five paleolakes of variable extension and duration. These paleolakes have been further described by Feibel (1997) from old to younger as paleolakes Lonyumun, Lokochot, Lokeridede, Lorenyang and Silbo (Fig. 15). Whether a single, continuous long-lived lake or several shorter-lived lakes occupied the Turkana Depression is of particular paleoenvironmental relevance, among others to characterize water resources for aquatic and terrestrial biota, including hominins, to understand the role of the various causal factors of environmental change, and to explain biogeographical dynamics, including both intra- and inter-basinal dispersal. The here reported sedimentological data (Fig. 13) reveal that continuous lacustrine conditions occurred in the North Lake Basin between ~4.00 and ~3.30 Ma and between ~2.30 and ~1.25 Ma, even if lake extent varied through time. The repeated occurrence of microbial constructions in strata of the Nachukui Fm between ~1.25 and ~0.75 Ma (Harris *et al.*, 1988) also supports recurrent lacustrine conditions in the area during that period. Our data and interpretations hence suggest that the basin, was occupied by a several tens of km-wide lake during at least ~70% of the period between ~4.00 and ~0.75 Ma. No direct evidence of lacustrine conditions is identified between ~3.30 Ma and ~2.30 Ma, which does not automatically imply the absence of a lake, however. Major incised valleys and canyons, which traditionally result from an important base level fall (Posamentier, 2001), are absent in this interval. This absence may support the existence of a lake although subaqueous offshore deposits are not observed in the available exposures. Hence, we are inclined to consider possible the view of a single long-lived large lake in the northern Turkana Depression between ~4.00 and ~0.75 Ma, even if this lake experienced important fluctuations and periods of particularly reduced spatial extent during, as examples, the climaxes of both the Lokochot (~3.60 to ~3.50 Ma) and Tulu Bor lake

lowstands (~2.70 to ~2.50 Ma). At that two periods, it is critical to estimate extent of paleolake Turkana along the North-South axis and some portions were possibly dry. Comparisons with previous studies reveal that three of the five palaeolake phases reconstructed by Brown and Feibel (1988, 1991), namely the Lonyumun, Lokochot and Lorenyang "paleolakes", coincide broadly with major highstands in our reconstruction. Exceptions concern the Lokeridede and Silbo phases. The Lokiredede paleolake is centred at ~2.50 Ma and considered to be brief (Feibel, 1997). Even if we did not observed deposits of this paleolake "phase", based on our results we suspect that it corresponds to a subordinate-order short-term transgression during the general Lomekwi Lowstand, comforting the presence of a lake during this important lake lowstand. Concerning the Silbo paleolake, which is centred around 1 Ma, we attribute this lake phase in the wake of the Nachukui Highstand centred at ~1.40 Ma, possibly representing a subordinate-order short-term transgression during the general regressive trend. However, our reconstruction differs from the previous ones by Brown and Feibel (1988, 1991) and Feibel (1988, 1997, 2011) mainly in the lowstand phases. These authors considered lowstand phases to be characterized by the absence of lacustrine conditions in the basin, whereas we interpreted basinward sections to regularly contain lacustrine deposits albeit that the spatial extent of aquatic environments was strongly reduced during these phases. In conclusion, a (quasi) long-lived, i.e., proto-lake Turkana, may have persisted during most of the Plio-Pleistocene in the Turkana Depression. It likely constituted an aridity refugium for terrestrial fauna, including hominins, as previously suggested by Joordens *et al.*, (2011), in contrast with some other lakes of the East African Rift. This long-lived lake may have limited east-west terrestrial connections for most of the time, and dampened speciation and radiation between western and eastern rift sides, in turn favoring north-south connections.

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5.4. Paleohydrography and aquatic community change

How aquatic environments in the Turkana Depression were connected to those in other ecoregions over time has important implications on aquatic biota and faunal turnover. Although much is still unresolved, especially as to the drivers of faunal turnover, we provide preliminary considerations on historical biogeography.

During the Lonyumun Highstand (~4.00-3.60 Ma), lacustrine environments covered most of the Turkana Depression. The paleolake was fresh and open, with an outflow to the Indian Ocean that drained the depression in the southeast (Cerling, 1979; Brown and Feibel, 1988; Bruhn *et al.*, 2011). The period coincided with persistent humid conditions in West Turkana, evidenced by an extensive woody cover, and the rivers discharging to the Turkana Depression likely ensured overall good hydrographic connectivity, both upstream and downstream of the paleolake. The mollusks that colonized the Turkana Depression during the Lonyumun Highstand support widespread faunal exchange with other regions in Tanzania, Kenya and Uganda, which were inhabited by lineages ancestral to the current Nilotic fauna (e.g. Van Bocxlaer *et al.*, 2008; Van Bocxlaer 2011; Schultheiß *et al.*, 2014; Ortiz-Sepulveda *et al.*, 2020). Some fish taxa, e.g. *Semlikiichthys*, also support widespread faunal exchange within a Nile-linked system (Stewart 2003) suggesting important rivers activity in the catchment of the paleolake and a hydrographic connectivity with both Nile-linked and central African river systems.

After the comparatively short-lived Lokochot Lowstand, during which the lake was endorheic, water levels rose to bring on the Lokochot Highstand (~3.50-3.30 Ma). During this highstand the outflow towards the Indian Ocean was likely re-established to conditions during the Lonyumun Highstand (Bruhn *et al.*, 2011), although the paleolake was slightly less extensive. A rainforest expansion towards the Turkana Depression was claimed by Williamson (1985) for this timeframe based on remains of the central African rainforest tree *Antrocaryon*

from sediments of the Usno Fm in the Omo Valley. This fossil evidence suggests substantial renewed river activity in the catchment of the paleolake and hydrographic connectivity with central African river systems. In terms of aquatic communities, information is more limited. Williamson (1985) also reported that the freshwater gastropod *Potadoma*, which is currently restricted to central and West African rainforest steams (Brown, 1994), appeared simultaneously with *Anthrocaryon* in the Turkana Depression. However, the geochemical correlation for this interpretation has been refuted (Van Bocxlaer *et al.*, 2008), and given that freshwater biota for the Lokochot Highstand remain poorly documented, further work is required to infer precise hydrographic connectivity and faunal affinities at that time.

After the Lokochot Highstand, a long-term progressive regression occurred (~3.30-2.70 Ma), marking a period of negative water budgets even if unfrequent second-order short-term transgressions likely occurred during the general regression. As a result, aquatic biota and ecosystems in the Turkana Basin were severely disrupted, but how exactly this disruption occurred and at which time exactly is currently unclear. Around 2.70-2.50 Ma, a colonization by a new assemblage of freshwater molluscs known as the Suregei Isolate fauna (Williamson, 1981; Van Bocxlaer *et al*., 2008) occurred coeval with a brief lake level rise during the Lomekwi Lowstand, i.e. 'the Lokeridede paleolake phase' *sensu* Feibel (1997). The invading taxa are known from fossil deposits elsewhere in Africa, notably from the Albertine Basin (Van Damme and Pickford, 1999, 2003, 2010; Van Bocxlaer *et al.*, 2008). Fossil evidence also suggests the invasion by freshwater sponges of the genus *Potamophloios* (Van Bocxlaer *et al.,* 2008), which at present only occurs in the Congo Basin and Zambia (Manconi and Pronzato, 2002). Hence, all evidence indicates that some rivers that drain into the Turkana Depression became hydrographically connected to regions more to the southwest, i.e. the Congo River catchment. Modifications of river catchments that fed the Turkana Depression further south, such as that of the Turkwel and/or Kerio River, have perhaps caused such faunal exchange.

The Suregei Isolate fauna disappeared from the Turkana Depression before the subsequent high lake level phase, the Lorenyang Highstand, which started ~2.10 Ma ago.

During the Lorenyang Highstand, the paleolake was open, with an outflow to the Mediterranean Sea according to Bruhn *et al.* (2011). The period directly associated to the Lorenyang Highstand marks the arrival of a new assemblage of freshwater biota, which persists up to ~1.40 Ma in the Turkana Depression, suggesting that it persisted during the Kaitio, the Natoo 1 and 2 Lowstands. This fauna contains endemic mollusks (Williamson, 1981; Van Bocxlaer *et al.*, 2008; Van Bocxlaer and Van Damme, 2009; Van Bocxlaer, 2011), which are marked by evolutionary change, and the freshwater stingray *Dasyatis* (Feibel, 1993). The mollusks associated with this phase differ from the previously mentioned Suregei Isolate fauna, implying that an ecological calamity caused extinction prior to recolonization. This phase of recolonization reflects the hydrographic reconnection of the Turkana Depression with surrounding drainage system. Given the indicated humidification in the Omo River catchment and simultaneous aridification at West Turkana, aquatic mollusks likely re-invaded the paleolake via hydrographic reconnections north of the Turkana Depression.

The final wave of freshwater faunal invasion recorded in the Omo Group is known as the Guomde Isolate (Williamson 1981; Van Bocxlaer *et al.*, 2008), occurred at ~1.10 Ma. This fauna has been recovered in the Koobi Fora, Nachukui, and Shungura Fms between ~1.10 and 0.75 Ma. The diversity of this fauna is lower than of that associated with the Lorenyang Highstand, but taxa sensitive to oxygen depletion and elevated water conductivity are present, suggesting a substantial period of stable conditions for freshwater life, and some morphological changes over time are observed in the fauna (Van Bocxlaer, 2011). This reveals that freshwater conditions were eventually re-established, although water was considered brackish during the Nachukui Highstand, at ~1.40 Ma (Cerling, 1979). The hydrographic

connections that have led to this invasion were probably similar in nature to those that enabled faunal dispersal during the Lorenyang Highstand, but they were probably less elaborate.

In conclusion, even though lacustrine conditions persisted for long periods in the northern Turkana Basin between ~4.00 and 0.75 Ma, the aquatic fauna of the northern Turkana Depression records multiple events of complete faunal turnover, notably, but not exclusively in freshwater mollusks (Van Bocxlaer *et al.*, 2008). Further investigations are required to understand (1) how these ecological calamities were brought on if not by desiccation; (2) to what extent aquatic and terrestrial taxa were similarly or differently affected by these calamities, and (3) how the Turkana Depression was hydrographically connected to other catchments in the East African Rift over time.

5.5. What was the impact of tectonism and volcanism on paleolake fluctuations?

Drivers of water level fluctuations in the Turkana Depression have been another matter of debate for decades. Some authors proposed that major lacustrine highstand in this basin derive from regional or global climate conditions (Trauth *et al.*, 2005, 2007; Maslin and Trauth, 2009; Maslin *et al*., 2014). Others (Feibel, 1988; Brown and Feibel, 1991; Bruhn *et al.*, 2011; Lupien *et al.*, 2018; Boes *et al.*, 2018) rather propose that palaeolake level fluctuations are primarily controlled by local tectonic or volcanic events, with a minor overprint of global to regional climate patterns. An emblematic example used to support this latter statement is the Lorenyang lake phase (i.e., highstand centred at ~2.10-1.90 Ma) which is attributed to the Lenderit Basalts flooding volcanic event that would have blocked the lake outlet of the proto-lake Turkana (Bruhn *et al.*, 2011), raising lake level as a consequence. Here, we demonstrate that the Lorenyang Highstand coincides with the largest negative shift in $\delta^{13}C$ isotopic values in the Omo Valley, revealing a coeval major increase of precipitation on the Ethiopian Highlands. Hence, whether or not volcanism have raised the paleolake outlet and

consequently increased the maximum potential available space to be filled by water, expansion of Lake Lorenyang and infilling by water of the newly-formed available space resulted from increased volume of water entering the basin due to wetter conditions onto the Ethiopian Dome. As such, the Lorenyang Highstand has been climatically-induced. Similarly, we reveal that the other high-amplitude paleolake fluctuations between ~4.00 and ~0.75 Ma were controlled by climate, as major lake highstands are associated with increased precipitations onto the Ethiopian Dome. Finally, we propose that tectonism and volcanisms had only a minor influence on proto-lake Turkana fluctuations

6. CONCLUSION

Combining reconstructions of sedimentary dynamics, paleolandscapes and paleovegetation, we provide an updated synoptic review of paleoenvironmental change in the West Turkana area, between ~4.00 and ~0.75 Ma. Our main conclusions are:

- Facies and sequence analyses reveal that proto-lake Turkana experienced seven high-amplitude transgression-regression (T-R) cycles between ~4.00 and ~0.75 Ma;which are, in places, superimposed by lower amplitude T-R cycles. Four major lake highstands are recognized and referred to as Lonyumum (~4.00-3.60 Ma), Lokochot (~3.50-3.40 Ma), Lorenyang (~2.10-1.90 Ma) and Nachukui (~1.40 Ma) during which proto-lake Turkana may have outflowed first toward the Indian Ocean and after ~2.20/2.00 Ma toward the Mediterranean Sea. These four major lake highstands are associated with two shorter-duration and less extensive lake highstands referred to as Katio (~1.75-1.50 Ma) and Natoo (~1.44-1.43 Ma). In between lake highstands, successive major lake lowstands occurred during which proto-lake Turkana was endoreic. An important short-term lake lowstand referred to as the

Lokochot Lowstand (~3.60-3.50 Ma) is identified at the end of the Lonyumun Highstand. The long-term Lomekwi Lowstand reaches a minimum between ~2.70 and ~2.50 Ma while three shorter-term lake lowstands referred to as Kaitio, Natoo 1 and Natoo 2 characterize the period between ~1.90 and ~1.45 Ma.

- The $\delta^{13}C$ values in soil carbonates allowed us to compare changes in sedimentation and the woody cover in both the West Turkana and the Omo Valley areas. This comparison reveals that the evolution of rainfall over the Ethiopian Dome and the drainage basin of the Omo River controlled high amplitude fluctuations of proto-lake Turkana during the Plio-Pleistocene period. In contrast, rainfall fluctuations over the West Turkana area did not significantly influence proto-lake Turkana water-level. Moreover, our study demonstrates that rainfall evolution differed continuously between the two areas and suggests that West Turkana climate is strongly influenced by local effects. Our results thus suggest that terrestrial faunas occupying West Turkana may have experienced significantly different climate conditions than those elsewhere in the Turkana Depression.

- Sedimentary dynamics in West Turkana show alternations between river- or wave-dominated lake margins over time. These changes are attributed to important variation in the sediment supply coming from the western rift shoulder that led to the alternations between typical "accommodation-" and "supply-"dominated systems through time. Pulses of activity of the western border fault (i.e., MRLF) repeatedly formed high relief, which caused repeated periods of increased denudation and sediment supplied to the basin. These high-relief periods are estimated to have occurred between ~3.4 and ~2.5 and after ~1.42 Ma.

- Climate processes that influenced Ethiopian Highlands, and significantly influenced the high-amplitude fluctuations of proto-lake Turkana likely derive from

the interaction of global climate events and the direct impact of insolation through precession cycles modulated by eccentricity variations.

- Whereas the reconstructed lacustrine highstands overall coincide well with previously recognized paleolake phases (Brown and Feibel, 1988, 1991, Feibel 1997, 2011), our reconstruction differs substantially during lake lowstands. Whereas these phases were previously reconstructed to have been river-dominated, our sediment analyses suggest that lacustrine conditions occurred for at least 70 % of the time. Considering all evidence, we interpret longer-lasting lake phases and potentially even a single, continuous paleolake between 4.00 and 0.75 Ma. Aquatic faunas display several major turnover events indicating that severe ecosystem calamities nevertheless occurred. Further studies are required to understand the underlying causes, and how faunal and environmental changes are linked in the Turkana Depression.

- Finally, our study suggests that tectonism, volcanism and climate interacted in intricate ways. Tectonism and volcanism have had an important effect on sedimentation modes; however, their impact on high-amplitude proto-lake Turkana water fluctuations and the establishment of the successive major lake highstands and lowstands appears to have been limited.

Since the paper by Harris *et al.*, (1988), this contribution is the first to present a synthetic investigation of the Nachukui Fm. Our study shows that such comprehensive studies can provide a basis for robust large-scale paleoenvironmental reconstructions and information on paleolake fluctuations, which complement detailed site by site examinations. We propose that similar approaches could be developed on other Plio-Pleistocene deposits, especially, but not exclusively, for other geographically extensive formations with hominin sites. Finally, the

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next step is to routinely integrate this approach in workflows adopted by paleontological and archeological research programss.

The following are the supplementary data related to this article.

Table S1. Geographic coordinates of measured sections.

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Conflict of Interest Statement

Authors of the manuscript entitled: "Plio-Pleistocene sedimentation in West Turkana (Turkana Depression, Kenya, East African Rift System): paleolake fluctuations, paleolandscapes and controlling factors" claims that they have NO conflicts of interest to declare.

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Fig. 1. Structural and geological background of the Turkana Depression (modified from Ragon *et al*., 2019). (A) Location of the Turkana Depression within the East African Rift System (EARS). (B) Simplified geology and main structures in the Turkana Depression. The background topography is derived from SRTM1. (C) Large-scale transects showing graben and half-graben architectures in the (a) northern and the (b) southern Turkana Depression (Africa Oil Report, 2016).

Fig. 2. Simplified geological map of the West Turkana area delineating members that composed the Nachukui Fm (modified from Harris *et al*., 1988). Distribution of faults originates from gravity map (unpublished data, AfricaOil). Georeferencing of sections is given in supplementary data (Table S1).

Fig. 3. Synthetic lithostratigraphic chart of the northern Turkana Depression and major tectonic events (modified from Ragon *et al*., 2019).

Fig. 4. Measured sections 1 to 18 (see locations in Figure 2). Lithologies and sedimentary structures allowed the interpretation of six depositional environments (Table 1). Based on the interpretation of depositional environments, high- and low-frequency T-R cycles are delineated. Available absolute ages are located along sections (see Table 2 for details). Major tuff-based and onfield visual correlations are indicated.

Fig. 5. Macroscopic facies observed in offshore (FA1) and transitional (FA2) deposits. A and B) Laminated mudstones (section 10). C) Diatomite bed intercalated in mudstones (section 16). D) Erosionally-based massive sandstone bed embedded in mudstones (section 12) interpreted as a turbidite deposited in offshore domain. E) Sandstone bed showing an erosional basal boundary overlain by planar laminations (upper flow regime) grading to wave ripple laminations (lower flow regime) recording temporal decrease of an oscillatory current (section 8). It is interpreted as a tempestite bed deposited in a transitional depositional

environment. F) Bedding plane view of cogenetic interferent wave ripples at the top of the tempestite bed (note the polygonal pattern of ripple crests that reflect multidirectional oscillatory currents), horizontal burrows reflect post-storm bioturbation subsequent to the deposition of sand (section 8). G) Centimetric packstone beds embedded in mudstones (section 5) interpreted as distal storm deposits; they reveal a transitional depositional environment. H) Top surface of the shelly bed (section 5) showing transported gastropod shells and internal molds.

Fig. 6. Macroscopic facies observed in wave-dominated nearshore (FA3) deposits. A) Well-sorted sandstones showing alternations between planar laminations and wave ripples, in places, intervals rich in heavy minerals are observed. They reflect sedimentation in a shoreface depositional sub-environment (section 12). B) Alternations between sandstone intervals and packstones attributed to shoreface deposits (section 12). C) bedding plane view of wave ripple: note the N-S oriented crests revealing E-W oriented paleowave displacement (section 4). D) Hummocky-cross stratified bed (section 3). E) Close-up view of packstones made of mollusk shells (*Melanoides*) in beach deposits (section 9). F) Packstones made of mollusk shells (*Bellamya*) in beach deposits (section 12). G) Packstones made of low-angle oblique stratifications that reflect successive beach faces. Low-angle stratifications suggest a relative dissipative beach (section 9). H) Moderate-angle oblique stratifications in packstones interpreted as a reflective beach (section 3). I) Bioclastic sandstones characterized by abundant bioturbations in the form of vertical burrows or root tubes. Several generations of bioturbated horizons connect to different levels that reflect successive paleosurfaces. This facies reflects a low-energy coastal swamp.

Fig. 7. Macroscopic facies observed in wave- (FA3) and river-dominated (FA4) nearshore deposits. A) Pluri-meter interval including sandstones and packstones in landward (West) dipping master bed overlain by a metric interval made of sandstones and packstones showing

low-angle lakeward (East) dipping master beds (section 12). Together, sediments reflect retrograding washover fan deposits topped by a prograding beach (see details in Nutz *et al*., 2017). B and C) Fine to medium matrix-supported conglomerates. Clasts are frequently coated by carbonates. Matrix is made of gastropod shells or shell fragments and sandstones (section 16). These deposits reflect river-dominated nearshore environment. D) M-scale oncolite-rich bed (FA4) embedded in river deposits (FA5-b). E) Interval showing matrix-supported oncolites in the lowermost part grading upward to clast-supported oncolites and then to domal stromatolitic constructions interpreted as the result of the interruption of river influence in the nearshore domain. F and G) Channelized sandstone beds embedded in mudstones. They are interpreted as mouth channels and lobes in river-dominated nearshore.

Fig. 8. Macroscopic facies observed in fluvial (FA5) deposits. A and B) Alternations of coarse sandstones to medium conglomerates in lenticular bodies and sandy mudstones to muddy sandstones. They reveal alternations of wide and thin channels and overbank deposits in an alluvial depositional environment. C) Bioturbated sandy mudstones and muddy sandstones. Termite nest are very well-preserved. Sediments reflect overbank deposits in alluvial plain (Table 2). D) Through cross stratifications in coarse sandstones revealing dunes and bars in alluvial channels. E) Erosionally-based m-scale interval made of fining-up matrix-supported conglomerate interpreted as cohesive debris flows deposit. F) Root marks in gravel bars suggesting recurrent interruptions of flows. G) Alternations of massive conglomerates revealing debris flows and coarse sandstones from dilute flows. They indicate type-1 alluvial fan deposits (*sensu* Blair and McPherson, 1994). H) Crudely laminated coarse sandstones to pebbles interpreted as type-2 alluvial fan (*sensu* Blair and McPherson, 1994).

Fig. 9. Large-scale sedimentary architectures of sequences 1 and 2. A) Thick interval of offshore lacustrine mudstones (Kataboi laga). B) Transition from offshore mudstones to alluvial deposits corresponding to the regressive part of S2. C) Close-up view on the transition

between offshore lake and alluvial plain deposits that expresses the emersion in the upper part of S2.

Fig. 10. Large-scale sedimentary architectures of sequences 2 and 3. A) Transition from offshore to shoreface and then alluvial deposits forming the regressive part of S2. B) Close-up view on SB3 that is amalgamated with the transgressive surface. C) Incised valley affecting offshore mudstones and filled by alluvial deposits in the upper most part of S2. Above, a beach interval marks the subsequent transgression associated with S3. D) Close-up view on the contact between offshore mudstones and alluvial sandstones to conglomerates. Desiccation cracks reveals emersion of the offshore mudstones before alluvial sedimentation.

Fig. 11. Large-scale sedimentary architectures of sequences 3 and 4. A) Alluvial deposits in the uppermost part of S2 overlain by beach and then offshore deposits marking the transgressive part of S3. B) Offshore mudstones marking the maximum flooding interval (MFI) of S3. C) Alluvial plain deposits corresponding to the regressive part of S3. D) Alluvial plain deposits in S4 that mark a renewed transgressive trend above the regressive part of S3 attributed to the lower part of S4. E) Beaches embedded in alluvial plain deposits evidencing the proximity of a lake. F) Offshore mudstones marking the maximum flooding interval (MFI) of S4 (Kalochoria laga). G) Shoreface deposits intercalated between alluvial sediments. This interval shows a transgression followed by a regression of paleolake Turkana that form S4 in the proximal domain (upstream northern branch of Kokiselei laga).

Fig. 12. Large-scale sedimentary architectures of sequences 5, 6 and 7. A) Coastal deposits in the uppermost part of S4 overlain by transitional deposits marking the transgression that forms the lower part of S5 (southern branch of Kokiselei laga). B) Coastal and alluvial deposits in the uppermost part of S6 overlain by offshore lake deposits marking the transgression that forms the lower part of S7 (Nachukui laga). C) Offshore mudstones

marking the maximum flooding interval (MFI) of S7 (Nachukui laga). D) River-nearshore deposits in alluvial fan conglomerates in the upper part of S7. E) Close-up view on oncolites comprised in the river-nearshore deposits. F) Alluvial plain interval in the upper part of S7 (Nariokotome laga).

Fig. 13. Interpretrative lithostratigraphic chart of the Plio-Quaternary succession (~4-0.75 Ma) in the West Turkana area. The chart is West-East oriented; successive lake highstands and lake lowstands are represented and a reconstruction of the paleolake Turkana extent through time is proposed. Periods coeval with the development of type-1 or type-2 sedimentary systems are delineated. Note only the high-amplitude water level fluctuations are presented. As this chart is synthetic, the lateral extent of the different depositional systems are approximate, potentially varying laterally from South to North.

Fig. 14. Conceptual depositional models representing ideal type-1 and type-2 sedimentary systems that alternatively characterized sedimentation in the West Turkana area between 4 and 1.25 Ma. Transition from one system to the other is due to variable sediment supply coming from the rift shoulder through time and reflects the transition from accommodation-dominated to supply-dominated systems and conversely (see text for details). Modern analogues are proposed for each model suggesting that they represent depositional systems observed in modern extensional basins.

Fig. 15. Comparison of lake level fluctuations and sedimentary systems (details Fig. 14) evolutions with accommodation evolution, woody cover, eccentricity (Laskar *et al.*, 2004) and previously published paleoclimate reconstructions. Note that only major tuff layers are represented.

Fig. 16. Paleolandscapes at key periods of the west Turkana evolution between ~4 and ~1.25 Ma. Types of sedimentary systems (see details Fig. 14), proportion of woody cover and

paleoclimate in both the West Turkana and the Omo River catchment areas are indicated for each period. Main carbonate producer is specified at each period.

Table 1. Facies description and interpretation.

Code	Lithologies	Texture, structures and bedding characteristics	Depositional processes	Depositional environments	
FA1	Mudstones	blackish, brownish or greenish; fine silt to clay; dm-scale draping beds; mm- to cm-scale horizontal laminations; common mm- to cm-scale low-angle oblique laminations; common dm-scale diatomite beds; fishbones; occasional dm-scale massive sandstone beds with erosional bases in places feeding injectites	suspension-fallout deposition; common low-density gravity flows and bottom currents; occasional high-density gavity flows; river-derived hyperpycnal flows; below the limit of fair-weather and storm-wave bases	Offshore lake (Scholle, 1971; Boulesteix et al., 2019)	
FA2	Mudstones, sandstones bioclastic packstones	dm-scale mudstone beds alternating with dm- to m-scale erosionally-based sandstone beds and cm-scale erosionally-based packstone beds; sandstones beds are made of either tempestite sequence topped by interference ripples or hummocky-cross stratifications; packstone beds include mollusk shells and shell fragments	suspension-fallout deposition during normal sedimentation; high-density gravity flows associated with storm conditions; high-energy oscillatory currents attributed to storm; below the fair-weather wave base and above the storm wave base	Transitional environment (Coe et al., 2003)	
FA3-a	Well-sorted medium to coarse sandstones, bioclastic (shells and shell fragments) sandstones, packstones and rudstones	wave ripples; planar laminations; occasional cm- to dm-through cross laminations; occasional heavy-mineral concentrations; occasional flat pebbles; common cm-scale vertical burrows	oscillatory currents; tractional currents in lower flow regime; occasional sorting of heavy minerals in the build up or surf zones; bioturbation	Shoreface (Clifton et al., 1971)	Wave-dominated nearshore (Nutz et al., 2017)
FA3-b		low- to moderate-angle planar laminations; occasional cm-scale vertical burrows	tractional currents in upper flow regime attributed to wave run-up in the upper swash zone; accretion of the beach face (Clifton 1971; Massari and Parea 1988)	Dissipative and reflective beaches; Foreshore (Clifton et al., 1971)	
		highly-bioturbated intervals with poorly visible primary structures separated by paleosurfaces	low-energy vegetated coast	Coastal swamp; Foreshore	
FA3-c		high-angle (25-30°) oblique stratifications dipping landward in 1-3 m-thick sediment bodies; frequent shells or shell fragments in updip cross-laminations; occasional wave ripples downdip cross-laminations; vertical burrows and rizoliths in the uppermost part	landward migration; avalanching; occasional oscillatory curents attributed to waves; soil develoment onto the top surface	Washover fans and sheets; Backshore (Neal et al., 2003; Baumann et al., 2017)	
FA4	Matrix-supported conglomerates	crude planar laminations, occasionally massive; matrix is made of poorly-sorted sandstones and bioclastic sandstones with common wave ripples and shell placers; frequent oncolites and oncolite beds	hyperconcentrated and/or debris flows; recurrent oscillatory currents (wave reworking); sorting of shells or shell fragments; frequent microbial crusts coating pebbles; frequent continuous microbial (stro-	Coastal alluvial fan (Bouton et al., 2016)	River-dominated nearshore

			matolites) beds intercalated in conglomerates		
	Well-sorted medium sandstones	dm- to m-scale tabular or lenticular beds interbedded with mudstones interval; planar laminations; rare climbing ripples and dm-scale cross-laminations; occasional load structures	turbulent density flows (Talling et al., 2012) associated with occasional dilute flows separated by periods of mud settling	River mouth complex (Schomaker et al., 2010; Nutz et al., 2015b)	
FA5-a	Poorly-sorted mudstones, sandstones, gravels and conglomerates	massive to horizontally-laminated mudstone beds; occasional fine sandstone beds with current ripples; frequent fossil termite nests; common burrows and rhizoliths; hundreds m-wide and m-thick lenticular erosive sediment bodies are embbedded in mudstone beds; fine to medium conglomerates and coarse sandstones; dm-scale planar cross-stratifications prevail; occasional dm-scale fining-up intervals with muddy and sandy matrix	suspension-fallout of muds alternating with tractional currents in lower flow regime attributed to dilute flows; alternations of overbank sand sheet deposits and low-energy fine sedimentation settling during maximum flooding; pedogenic faunal bioturbation and palaeosoil developments between flood events; lenticular sediment bodies are channels made of dunes and bars due to tractional currents in lower flow regime and debrites from occasional cohesive debris flows	Alluvial plain (Miall, 1996; Bridge, 2003)	River
FA5-b	Matrix-supported coarse conglomerates	heterometric pebbles and occasionally boulders supported by sandy matrix; commonly oriented, in places imbricated or vertical; frequent dm-scale fining-up interval with pebbles supported by a muddy sand matrix; occasional rizoliths	unconfined cohesive debris flows (Blair and McPherson, 1994, 2009); occasional dilute flows in upper flow regime; frequent soil develoments during periods of non-activity	Type-1 alluvial fan (Blair and McPherson, 1994, 2009)	
	Matrix-supported fine to medium conglomerates	dm- to m-scale coarse sandstone beds interbedded with cm- to dm-scale poorly-sorted fine conglomeratic beds; angular to sub-angular pebbles	unconfined hyperconcentrated flows (sheetflood); sediment-laden turbulent unconfined flows	Type-2 alluvial fan (Blair and McPherson, 1994, 2009)	

Tuffs identifed by chemical analysis

Section	Sample	n	Fe_2O_3	CaO	K_2O	Mn	Ti	Name	Age (Ma)	References
3	Tuff_14.1.2	14	4.39	0.17	0.59	1422	1925	K82-742	3.6 ± 0.01	2
3	Tuff_004	7	2.97	0.40	1.01	569	1477	Loruth Tuff	3.97 ± 0.05 > T > 3.6 ± 0.05	1
5	Tuff_3.23	14	3.13	0.19	2.20	1032	1449	K85-2860	3.97 ± 0.05 > T > 3.6 ± 0.05	1
6	Tuff_202	16	1.53	0.34	2.55	682	1124	Tulu Bor β	3.44 ± 0.02	1, 6
7	Tuff_3.24	9	2.66	0.20	1.21	774	980	Moiti Tuff	3.97 ± 0.03	1, 5
7	Tuff_4.24	14	3.19	0.18	3.21	1069	1381	K85-2860	3.97 ± 0.05 > T >	1

									3.6 ± 0.05	
7	Tuff_5.24	13	1.61	0.29	1.47	380	900	Tulu Bor β	3.44 ± 0.02	1, 5
8	Tuff_016	14	1.60	0.32	1.11	352	985	Tulu Bor β	3.44 ± 0.02	1, 5
9	Tuff_018	13	3.21	0.16	0.83	1187	1519	Lokalalei Tuff	2.53 ± 0.02	1; 5
9	Tuff_019	15	5.88	0.20	1.17	2710	2778	Kokiselei Tuff	2.40 ± 0.05	1, 7
10	Tuff_020	14	4.28	0.21	0.61	1348	1521	Nalukuwoi Tuff	2.40±0.05 > T > 2.33 ± 0.02	1
10	Tuff_021	12	6.08	0.24	0.33	1472	1130	K84-1849	2.33 ± 0.02 > T > 2.06 ± 0.02	1
10	Tuff_022	not analyzed						KBS Tuff	1.87 ± 0.02	1, 4, 6
11	Tuff_1.22	10	4.79	0.20	0.53	1484	1973	Tuff K82-740	2.33 ± 0.02 > T > 2.06 ± 0.02	1
12	Tuff_022	not analyzed						KBS Tuff	1.87 ± 0.02	1, 4, 6
13	Tuff_1.7	not analyzed						KBS Tuff	1.87 ± 0.02	1
13	Tuff_3.7	16	6.58	0.19	0.39	2281	2287	Etirr Tuff	1.41 ± 0.02	3
15	Tuff_4.20.2	5	5.03	0.23	0.79	1546	1811	Karari Grey Tuff	~1.43	3
16	Tuff_2.10	20	4.73	0.26	2.00	2032	3247	K83-1647	~1.41	3
17	Tuff_2.8	16	5.07	0.19	0.41	1753	1850	Lower Nariokotome Tuff	1.30 ± 0.03	1
18	Tuff_1.10.1	14	3.95	0.20	0.98	1018	1693	Middle Nariokotome Tuff	1.28 ± 0.03	1

(1) Harris et al., 1988; (2) Pickford et al., 1991; (3) Brown et al., 2006; (4) Lepre et al., 2011; (5) McDougall et al., 2012; (6) Harmand et al., 2015; (7) Boës et al, 2018

Table 2. Geochemical analyses for tephra sampled in the Nachukui Fm (all values are in wt%) and resulting identification for each tuff based on comparison with literature.

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